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GEORGIA INSTITUTE OF TECHNOLOGY
OFFICE OF CONTRACT ADMINISTRATION
SPONSORED PROJECT INITIATION

Date: 1/9/79

Project Title: A Seismic Spectral Discriminant For Reservoir Induced Earthquakes

Project No: G-35-645 Green card

Project Director: Dr. L. T. Long

Sponsor: U. S. Geological Survey; Menlo Park, CA 94025

Agreement Period: From 11/27/78 Until 11/26/79 (R&D Perf. Period)

Type Agreement: Contract No. 14-08-0001-17713

Amount: $19,526.00

Reports Required: Quarterly Management Reports; Semi-Annual Tech. Report; Final Tech. Rep

Sponsor Contact Person (s):

Technical Matters

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Defense Priority Rating: none

Assigned to: Geophysical Sciences (School/Laboratory)

COPIES TO:

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- Other

CA-3 (3/76)

GEORGIA INSTITUTE OF TECHNOLOGY
OFFICE OF CONTRACT ADMINISTRATION

SPONSORED PROJECT TERMINATION

Date: 8/12/80

Project Title: A Seismic Spectral Discriminant For Reservoir Induced Earthquakes

Project No: G-35-645

Project Director: Dr. L.T. Long

Sponsor: U.S. Geological Survey; Menlo Park, CA 94025

Effective Termination Date: 11/26/79

Clearance of Accounting Charges: 12/26/79 (for reporting)

Grant/Contract Closeout Actions Remaining:

- _ Final Invoice and Closing Documents
- _ Final Fiscal Report
- X Final Report of Inventions
- X Govt. Property Inventory & Related Certificate
- _ Classified Material Certificate
- _ Other ______________________

Assigned to: Geophysical Sciences (School/~~Laboratory~~)

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CA-4 (1/79)

GEORGIA INSTITUTE OF TECHNOLOGY
SCHOOL OF GEOPHYSICAL SCIENCES

Atlanta, Georgia 30332
(404) 894-2857

October 1, 1979

Dr. Jack F. Evernden, COR
U. S. Geological Survey
345 Middlefield Road
Menlo Park, California 94025

Subject:	Quarterly Management Report No. 1 covering period of 1 December 1978 to 28 February 1979
Date submitted:	September 15, 1979
Research Title:	A seismic spectral discriminant for reservoir induced earthquakes
Contract Number:	U. S. Geological Survey 14-08-0001-17713
Contractor:	Georgia Institute of Technology
Contract Period:	November 27, 1978 to November 26, 1979
Amount:	$19,526
Principal Investigator:	Leland Timothy Long

Dear Dr. Evernden:

During the first quarter of the project major efforts were directed first toward augmenting our spectral computation computer program to include a segmented spectral averaging technique which we hope will reduce scatter at higher frequencies. In addition we evaluated the influence on spectral computation of magnitude distance instrument response and attenuation. This study was to allow us to be more specific in requests for data.

We also searched for possible sources of appropriate data.

1. <u>Major accomplishments</u>: We were able to show that spectral estimates at higher frequencies can be improved significantly with a segmented spectral computation program.

2. <u>Problems Encountered</u>: No significant management problems were encountered.

3. <u>Fiscal Status</u>: Out of $19,526 total funds available $3,528.95 were expended during the quarter. Available funds should be sufficient to complete this study by December 1979.

4. <u>Action Required by USGS</u>: No action is requested.

5. Future Plans: No change in the program of study is planned.

6. Inventory of Property Acquired During Report Period: No property was acquired during the report period.

Respectfully submitted,

Leland T. Long
Associate Professor

LTL/dp

GEORGIA INSTITUTE OF TECHNOLOGY
SCHOOL OF GEOPHYSICAL SCIENCES

Atlanta, Georgia 30332
(404) 894-2857

October 1, 1979

Dr. Jack F. Evernden, COR
U. S. Geological Survey
345 Middlefield Road
Menlo Park, California 94025

Subject:	Quarterly Management Report No. 2 covering period of 1 March 1979 to 31 May 1979
Date submitted:	September 15, 1979
Research Title:	A seismic spectral discriminant for reservoir induced earthquakes
Contract Number:	U. S. Geological Survey 14-08-0001-17713
Contractor:	Georgia Institute of Technology
Contract Period:	November 27, 1978 to November 26, 1979
Amount:	$19,526
Principal Investigator:	Leland Timothy Long

Dear Dr. Evernden:

During the second quarter we continued our evaluation of parameters such as instrument response and absorptive attenuation which determine our ability to calculate spectral slopes above the corner frequency. Our results have been disappointing since appropriate data are in a very limited range of magnitudes and distance. Only low-gain (less than 50,000) WWSSN stations can be used for M_L about 3.0 ± 0.3 at about 50 ± 10 km range. All other events will give questionable spectra at the higher frequencies. Typical telemetry systems require magnitudes of 2.0 ± 0.5 at 15 ± 5 km recorded at relatively low-gain (less than 100,000) in order to allow computation of reliable spectra. It appears that only high-dynamic range data or data recorded under unusual circumstances are appropriate for our studies. During the quarter we initiated our requests for data. The quarter also included travel to the Seismological Society of America meeting and an informal symposium on reservoir induced seismicity.

1. Major accomplishments: The only real accomplishment was the definition of recording systems and events appropriate for spectral slope computation.

2. Problems Encountered: The limited range of data appropriate for spectral computation was a big disappointment. No significant management problems were encountered.

3. Fiscal Status: Out of $19,526 total funds available, $7,440.59 were expended by the end of the second quarter. Available funds should be sufficient to complete this study by December 1979.

4. Action Required by USGS: No action is requested.

5. Future Plans: We have decided to obtain and perform additional spectral computations on local southeastern U.S. data because other data are sparse. No other change in the program of study is planned.

6. Inventory of Property Acquired During Report Period: No property was acquired during the report period.

Respectfully submitted,

Leland T. Long
Associate Professor

LTL/dp

G-35-645

GEORGIA INSTITUTE OF TECHNOLOGY

SCHOOL OF GEOPHYSICAL SCIENCES

Atlanta, Georgia 30332
(404) 894-2857

October 1, 1979

Dr. Jack F. Evernden, COR
U. S. Geological Survey
345 Middlefield Road
Menlo Park, California 94025

Subject:	Quarterly Management Report No. 3 covering period of 1 June 1979 to 31 August 1979
Date submitted:	September 15, 1979
Research Title:	A seismic spectral discriminant for reservoir induced earthquakes
Contract Number:	U. S. Geological Survey 14-08-0001-17713
Contractor:	Georgia Institute of Technology
Contract Period:	November 27, 1978 to November 26, 1979
Amount:	$19,526
Principal Investigator:	Leland Timothy Long

Dear Dr. Evernden:

During the third quarter we continued to search for appropriate data. However, we have had very little success. While some data has been received we have not been able to identify data that are appropriate. Most people have been very cooperative. Only one organization has refused to allow examination of appropriate data. Few others have been understandably reluctant because of the large amount of data available to sift through and the time that might be involved. Any extension of this work should include travel monies to visit the laboratories where useful data might be found. It is becoming very clear that proper instrumentation in an area prior to reservoir impoundment is important. The existence of appropriate data is very limited. To help fill the data gap we have started additional analysis on events from Lake Sinclair and Clark Hill Reservoir area. We have also attempted to record some events at Montecello, S. C. While these will augment the data base they are still only from a restricted area.

1. Major Accomplishments: Although no great accomplishment was realized during the quarter, steady progress was made.

2. Problems Encountered: Still, no significant data has been obtained. No significant management problems were encountered.

3. Fiscal Status: Out of $19,526 total funds available, $16,310 were expended by the end of the third quarter. Available funds should be sufficient to complete this study by December 1979. Additional travel funds might be helpful.

4. Action Required by USGS: No action is requested.

5. Future Plans: In additional to placing more emphasis on southeastern U.S. reservoir related events we are considering using available data during the next quarter to evaluate Q in the southeastern U.S.

6. Inventory of Property Acquired During Report Period: No property was acquired during the report period.

Respectfully submitted,

Leland T. Long
Associate Professor

LTL/dp

Semi-annual Technical Report Number 1

A SEISMIC SPECTRAL DISCRIMINANT FOR RESERVOIR INDUCED EARTHQUAKES

Leland Timothy Long, Principal Investigator
and Greg Johnston
School of Geophysical Sciences
Georgia Institute of Technology

June 26, 1979

Sponsored by the U.S. Geological Survey
Contract No. 14-08-0001-17713

Semi-annual Technical Report Number 1

Contract Number: 14-08-0001-17713

Name of Contractor: Georgia Institute of Technology
School of Geophysical Sciences

Principal Investigator: Leland Timothy Long

Government Technical Officer: Jack F. Evernden

Title of Work: A Seismic Spectral Discriminant for Reservoir Induced Earthquakes

Effective Date of Contract: November 27, 1978

Contract Expiration Date: November 26, 1979

Amount of Contract: $19,526.00

Date Report Submitted: June 26, 1979

Semi-annual Progress Report #1

"A Seismic Spectral Discriminant for Reservoir Induced Earthquakes"

Leland Timothy Long and Greg Johnston

June 26, 1979

Project Summary

The specific problem addressed by this research is the identification of areas where reservoirs could induce significant seismic activity. An ω-cube high frequency spectral slope was found to be characteristic of earthquakes associated with two southeastern United States reservoirs which have induced seismic activity. In contrast, microearthquakes in the Folded Appalachians have an ω-square high-frequency spectral slope in areas where reservoirs have not induced earthquakes. Hence, in the southeastern United States the spectral slope may be a viable discriminant for identify areas susceptable to induced seismic activity. In this study the generality of the spectral discriminant is to be tested for earthquakes of other reservoir areas.

Fundamental to this study is an understanding of the theoretical basis for the discriminant. Source models have been reviewed and summarized as to their predicted high-frequency spectral slopes. The ω-cube high-frequency decay is typically associated with transsonic rupture velocities on existing plains of weakness. The ω-square high-frequency decay is typically associated with subsonic rupture velocities on fault plains which may shown premature arrest of slip.

Published reports of spectral slopes have been examined in the hope of discriminating ω-square or ω-cube areas. The remainder of this study will be concerned with testing the generality of the ω^{-3} high-frequency spectral characteristic in specific reservoir areas. Measured slopes will be compared to geology, regional tectonics, and spectra, when available, of near-by tectonic earthquakes.

The sparse availability of data of sufficient quality is expected to be the major difficulty to overcome in this study. Propagation effects and instrument response limit the data that can be used for high frequency spectral computation. In an exercise which involve computation of attenuation recording instrument saturation and frequency response the characteristics of data which may be successfully analyzed are identified. For WWSSN data recorded at a magnification of 50 k the high-frequency slope can be computed for typical magnitude 3 to 3.5 events recorded in the range of 40 to 180 km provided the Q value is greater than 800. In general, the success of the measurement of spectral slope (and hence of the discriminant for induced seismic activity) depends on the Q value for the data region if traditional and existing recording systems are to be used. Q values for the Georgia-South Carolina Piedmont area will be calculated. Using the method of spectral ratios, preliminary indications are that Q_p>600 for local events.

Possible reservoir data areas have been tabulated. Of these, only a few may have data which could be analyzed for the high-frequency spectral slope. Data from these areas are being requested.

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A SEISMIC SPECTRAL DISCRIMINANT FOR
RESERVOIR INDUCED EARTHQUAKES

I. Introduction

The initiation or enhancement of seismicity by the creation of some artificial lakes has been well documented. Earthquakes at a few of these reservoirs have been destructive, exceeding magnitude 6. However, reservoirs which have not exhibited induced seismicity are more numerous. The subject of this study is the evaluation of a criteria which may identify the possibility of induced seismicity at a given reservoir. The definition of the proposed criteria and the development of some of the techniques for its application to local earthquakes is the subject of this first report.

In the Southeast United States, several reservoirs in the Piedmont Province have induced seismic activity (Talwani; 1976, 1978). Earthquakes associated with these reservoirs show high-frequency displacement spectral slopes which decay as ω^{-3} past the corner frequency (Marion, 1977). In contrast, reservoirs in the Folded Appalachian Province have not generated noticeable activity and earthquakes in that province typically show ω^{-2} spectral decay. Therefore, the ω^{-3} high-frequency spectral slope appears characteristic of areas susceptible to induced seismicity in the Southeast United States and is the criteria proposed for use in identifying areas susceptible to induced seismicity.

In this study the generality of the ω^{-3} high frequency spectral slope is tested by spectral analysis of data from other reservoir areas. Unfortunately, data are available for only a few reservoir areas and can be difficult to obtain. Also, source, propagation and recording parameters can limit the computation of the high-frequency spectral slope. Seismic records from instruments with sufficiently well defined response at high frequencies to allow determination of the high frequency slope must be used. Smaller earthquakes may have corner frequencies above the reliable response of many seismic systems. Larger earthquakes with lower corner frequencies will be unsaturated only at more distant stations and the influence of attenuation may be uncertain. Also, many seismic writing systems run at such slow speed that the ability to follow the trace for spectra computation could be questionable. Hence, seismograms of earthquakes of a limited size recorded at restricted distances must be used in conjunction with a selected seismic recording system to allow the high-frequency seismic data analysis to be successful.

Since the ω^{-3} spectral decay in the Southeast United States can be explained in terms of an earthquake mechanism proposed for reservoir induced activity, the generalization to other areas would provide an explanation for why not all reservoirs induce seismic activity and may also allow prediction of the activity level to be expected for a particular reservoir.

II. Reservoir Seismicity

Perhaps the first recognized case of reservoir associated seismicity was at Lake Mead (Hoover Dam) in the late 1930's. Carder (1945) suggested that the water loads of Lake Mead reactivated the pre-existing faults in the area. Although a few other isolated cases of induced seismicity occurred before 1960, little attention was given to the phenomenon and it was thought that only small earthquakes caused by loading were associated with artificial lakes.

During the 1960's a series of events occurred that tended to support the decrease in strength of rocks with increased fluid pressure Hubbert and Rubey (1959) and the possibility that earthquakes might be triggered by artificially induced changes in the fluid pressure in the earth's crust. The apparent correlation between the rate of fluid injection at high pressure into the Rocky Mountain arsenal near Denver, Colorado and a subsequent series of seismic shocks in the same region is well documented (Evans, 1966). Also, the Rangely Oil Field, Colorado, microearthquakes were associated with secondary recovery water injection (Raleigh, 1976).

New or increased seismic activity has often been reported in reservoir areas. Damaging earthquakes occurred near large reservoirs in Kariba in Rhodesia (1963), at Kremasta in Greece (1966) and at Koyna in India (1967). These earthquakes caused tremendous destruction with the Koyna event the most significant; claiming 200 lives, injuring thousands and leaving more homeless. At Vajont Dam in Italy, in 1963 a major landslide and ensuing flood, related to and perhaps caused by induced seismicity, claimed about 2000 lives. Many more examples of induced seismic activity have been discovered and there are now seven accepted examples of sites where major seismic events have been associated with reservoirs. There are also many more (up to about 30) cases where minor induced seismicity has been detected (Milne, 1976). However, there are thousands of reservoirs where seismic activity has not been noted. This lack of activity, perhaps, is not surprising because only a very few reservoirs are instrumented to record local events both before and after the reservoir fills with water.

The stresses caused by loading and pore pressure in a reservoir area are comparitively small. The added stress due to water loads seldom exceeds 10 bars. Pore pressure increases caused by reservoirs could be a few tens of bars. The strength of crystalline rocks may be on the order of 1,000 bars. Since the stresses caused by impounded waters are small in comparison with the strength of crystalline rocks, it is assumed the rock masses involved were close to failure before filling of the reservoir commenced (Gupta and Rastogi, 1976) or perhaps the earthquakes occur along existing zones of weakness.

Although in particular cases either pore pressure or loading may be the dominant mechanism of reservoir induced seismicity, the presence or absence of seismicity is currently poorly understood and probably depends on a complex interaction of the reservoir with tectonic, geological, and hydrological factors.

III. Seismic Source and High-Frequency Spectral Slopes

The problem of inferring characteristics of the source mechanism of seismic events from the analysis of far-field seismic data has had much attention. In recent years, seismologists have attempted to fit observed body waves to various models of seismic source functions. Many investigators have obtained relationships between the character of the source-time displacement function and the far-field displacement spectra.

For most models, the displacement function for faulting with a step-like time function has, typically, a low frequency portion approximated by a constant displacement amplitude. The magnitude of the zero frequency portion is proportional to the seismic moment. The high-frequency trend is determined by the Fourier transform of the source time function and has an asymptotic spectral decay proportional to ω^{-n}

where n is found (Molnar et al., 1973) to be ≥ 1.5. The corner frequency is defined by the intersection of the two trends and behaves inversely with the pulse width of the displacement signal. The corner frequency decreases as the fault length increases. Conversely, a smaller pulse width is associated with smaller earthquakes and higher corner frequencies.

Figure 1 shows a theoretical displacement spectra compared with the displacement spectra of a typical Southeast United States earthquake of the Piedmont physiographic province. Observed spectra rarely have a smooth shape and normally show modulation. This modulation can be introduced at the source by a finite rupture propagation velocity or by multiple arrivals from waves traveling different propagation paths (Pilant and Knopoff, 1964).

The detailed shape of the seismic spectrum is controlled by many factors. The high-frequency behavior of the spectrum is determined by the manner in which the rupture area develops (Savage, 1966, 1972; Dahlen, 1974). The dimension of faulting controls the first P-and S-wave corner frequencies. The degree of seismic energy lost between source and station because of internal friction can effect the slope at higher frequencies (Tanis, 1973). The change of spectral content with earthquake magnitude in a given seismic area (Chouet et al., 1978) also modifies the spectrum. Larger earthquakes may have multiple ruptures and the interference of the multiple ruptures can affect the spectral slope. Consequently, many feel the waves generated by small earthquakes provide a more reliable determination of source parameters.

Most theoretical models of the seismic source mechanism for shallow focus events have utilized a fracture or dislocation model. Several

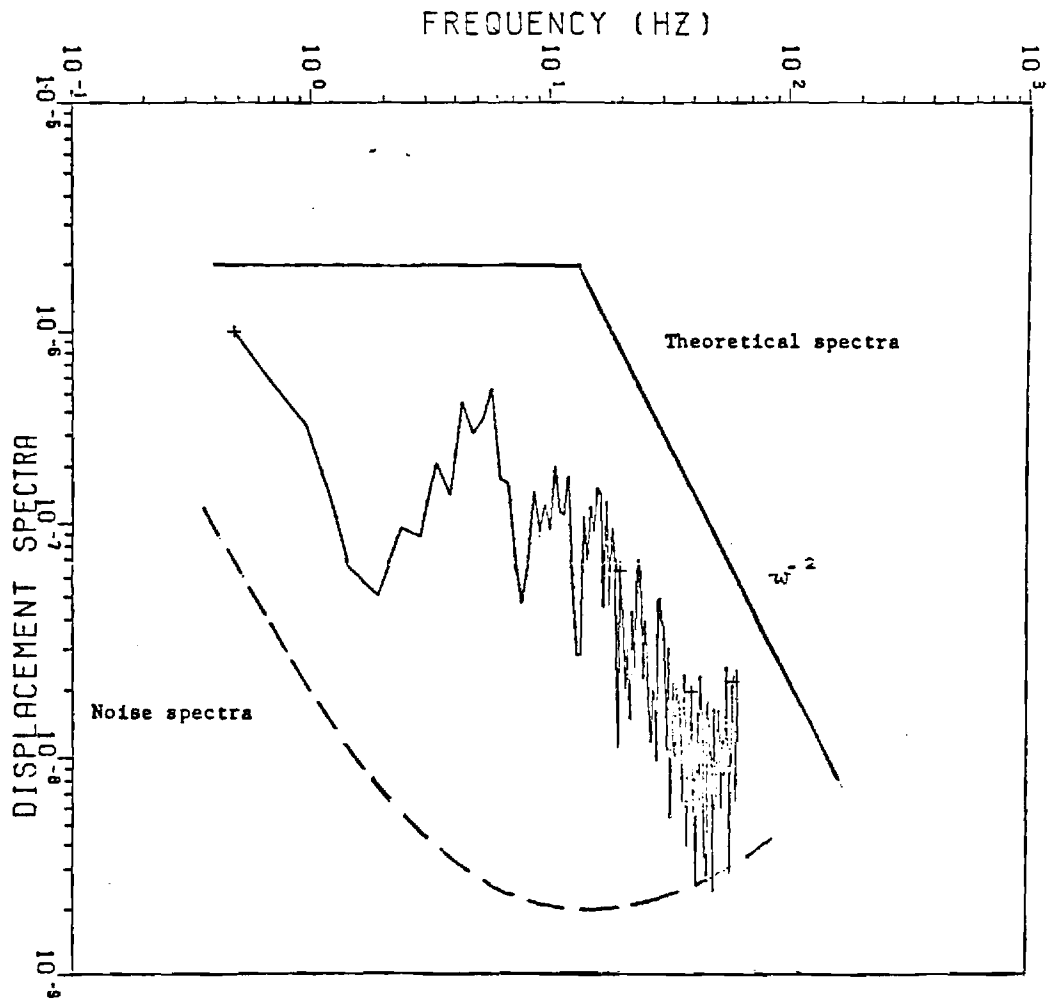

Figure 1. Theoretical displacement spectra (after Hanks and Wyss, 1972) compared to spectra of a southeast United States earthquake shown with typical noise spectra.

such models (Haskell, 1964, 1966; Savage, 1966, 1972; Aki, 1967; Brune, 1970, 1971; Molnar _et al_., 1973; Dahlen, 1974; Madariaga, 1976) have been constructed. An early paper which related seismic source parameters to the source spectrum was that of Aki (1967). He proposed two statistical models of seismic sources, an "ω-square" model (which decayed as ω^{-2} at high frequencies) and an "ω-cube" model (which decayed as ω^{-3} at high frequencies) after Haskell (1966).

Perhaps the most popular source model is that of Brune (1970) for S-wave displacement spectra. This model was extended to P waves by Hanks and Wyss (1972). Brune's model allows interpretation of easily measured features of seismic body-wave spectra in terms of a few gross physical parameters which characterize the source.

The Brune (1970) and Hanks and Wyss (1972) models assumed fault slippage on a circular fault which occurs simultaneously over the entire fault area, inferring infinite rupture speed for a finite time duration. In the far field this produces a displacement slope proportional to t exp $(-\alpha t)$. Where t is time and α is a model dependent constant. Brune's displacement function has a smooth spectrum with a spectral slope of ω^{-2} for frequencies higher than the corner frequency.

Savage (1972) objected to Brune's ω^{-2} model on the basis that realistic models of fault rupture which nucleate from a point and spread with subsonic velocity should produce an initially quadratic growth in the far field with a displacement slope proportional to t^2 exp $(-\alpha t)$. In the frequency domain, this implies a spectral amplitude which decays as ω^{-3} above the corner frequency (Bracewell, 1965).

This was verified by Molnar _et al_. (1973) who experimented with different "rupture velocities" and found the spectra decreased as ω^{-3} at high frequences. Dahlen (1974) obtained a ω^{-3} spectral decay by means of a three-dimensional kinematic faulting model which would arise if mechanical friction must be overcome to produce slippage on a fault plane.

For shallow earthquakes, the observed spectrum can be complicated by the effects of the fault length and width. As the length to width (L/W) of the fault area increases, Savage (1972) finds the width of the region between the rise time and fault-width corner frequency to increase. Geller (1976) found that large shallow earthquakes have generally L/W equal to 2. This gives higher corner frequencies for the large shallow earthquakes. In this model, Savage (1972) predicts the high frequency radiation past the corner frequency to decay also as ω^{-3}.

Other investigators such as Richards (1973), Sato and Hirasawa (1973) and Madariaga (1976) have outlined crack models for which the initial rupture contributes an ω^{-3} high frequency spectrum while a "stopping phase" caused by simultaneous cessation of fracture everywhere on the fault creates more high frequencies and causes an ω^{-2} to $\omega^{-2.5}$ high-frequency spectral decay. This agrees with the findings of Das (1976) that the high-frequency versus low-frequency content of the spectrum is also larger in a rupture with barriers than for a smooth rupture.

In terms of stress drop, Das and Aki (1977) determined that an earthquake with low average stress drop may generate relatively greater amounts of high-frequency waves than an earthquake with a high average stress drop. This tends to imply that a lower stress drop is associated with an ω^{-2} and a high stress drop with an ω^{-3} high frequency spectral slope.

IV. Observed Spectral Data

Marion (1977) calculated 165 body-wave displacement spectra from 93 microearthquakes in the Southeast United States. He found earthquakes from the Clark Hill and Jocasse reservoir areas, located in the Piedmont Province and considered to be examples of induced seismicity, to typically have a high stress drop and show an ω^{-3} amplitude decay at high frequencies. Conversely, earthquakes from near Maryville, Tennessee, located in the Folded Appalachian Province where reservoirs have not induced seismic activity, typically show an ω^{-2} to $\omega^{-2.5}$ high frequency decay.

The ω^{-3} high-frequency displacement spectral decay has been interpreted (Marion, 1977) as suggesting a transonic rupture (rupture velocity greater than S-wave velocity) on a fault surface with little or no frictional resistance. Guinn (1977) used focal mechanisms of Southeast United States earthquakes to determine a general tensional environment for the shallow Clark Hill and Jocasse reservoir events.

The ω^{-2} high-frequency displacement of the Maryville, Tennessee area earthquakes is described by Marion (1977) as being caused by a subsonic rupture (rupture velcoity less than S-wave velocity). This area is generally believed to be under a compressional stress environment.

Most published spectral data, while usually nonconclusive as to source properties and high frequency slopes can sometime give general ideas as to the type of source mechanism of their respective areas.

Table 1 lists observed published spectral data as to author, area, and spectral characteristics found.

Thatcher and Hanks (1973) studied the source characteristics of southern California earthquakes with local magnitudes M_L between 2 and 7. They observed a ω^{-2} high frequency decay for spectra at the closest stations but found an $\omega^{-1.5}$ decay for distant stations. Aki and Chouet (1975) calculated the average source spectrum for events recorded at Stone Canyon, California near the San Andreas fault. The spectra were observed to decay with a frequency roughly proportional to ω^{-2} beyond the corner frequency.

There have been no reports of induced seismicity in southern California. Oroville reservoir, in eastern California, has been the only California reservoir considered to have induced seismicity, although the case is not as clear as many others. Bufe et al. (1976) and Bell and Nur (1978) concluded normal faulting was the dominant faulting mechanism at Oroville. If southern California is under a general compressional stress environment as suggested by Hofmann (1973) perhaps this could explain why the high frequency waves of some southern California events were found to decay as ω^{-2} and why no induced reservoir seismicity has occurred.

Seismic data from the Lake Mead area were obtained for a data period of 1.5 years in 1972-1973 by Rogers and Lee (1976). Focal mechanisms showed fault-plane solutions that support right-lateral, strike-slip motions along nearly vertical faults. For some earthquakes, normal faulting was also possible. This analysis could tend to support a tensional stress environment for the Lake Mead area. Focal mechanisms of 207 small earthquakes were used by Wang et al.

Table 1. Observed published spectral data

Author	Area	Spectral Characteristics
Aki and Chouet (1975)	Stone Canyon, CA near San Andreas	ω^{-2} high frequency decay of coda waves
Bakun and Bufe (1975)	Bear Valley, CA	S-wave spectra $\omega^{-1.5}$ to ω^{-2} at > 2 Hz, fc for $1.1 \leq M \leq 2.2$ at >10-12 Hz
Bakun et al. (1976)	Central CA	frequency $\geq$10 Hz SH spectra decrease more rapidly than PZ spectra. Q_p = 175-250 Qs = 100-150, ω^{-2} to ω^{-4} high frequency decay
Bakun et al. (1978)	San Andreas Fault, CA	No slope given
Frasier and North (1978)	Rat Island, Southwest of Alaska	ω^{-3} high frequency spectral decay
Hanks and Wyss (1972)	CA, Iran, Turkey	Interpreted spectral data to decay as ω^{-2} at high frequencies
Johnson and McEvilly (1974)	Central CA	ω^{-2} to ω^{-3} high frequency decay
Peppin (1976)	Nevada Test Site	explosion data decay $\geq \omega^{-3}$ higher corner frequency than earthquakes
Peppin and Simila (1976)	Trans Sierra-Nevada CA-NEV	P and SV spectra show ω^{-2} or ω^{-3} high frequency decay $Q_s \geq 480$
Ryall et al. (1976)	Oroville Reservoir, CA	P and S-wave spectra ω^{-2} to ω^{-3} decay - Mag 3.0-4.3, corner frequencies 10-20 Hz
Tanis (1973)	Worldwide earthquakes	Coda displacement spectra slope not found
Thatcher and Hanks (1973)	Southern CA	ω^{-2} high frequency decay close in, $\omega^{-1.5}$ at distant stations
Wyss et al. (1971)	Aleutian Islands	Spectral data $0.5 \leq T \leq 33$ sec. ω^{-2} for earthquakes and explosions.

Wyss and Hanks (1972)	San Fernando, CA	S-wave spectra decay as $\omega^{-1.5}$ P-wave spectra decay as $\omega^{-1.8}$
Wyss and Shamey (1975)	Tonga Islands Kamchatka Islands	Long period spectra No slope found

(1976) to determine a predominently dip-slip mechanism of recent faulting near the Hsinfengkiong Reservoir, China.

Snow (1972) has theoretically shown that in a thrust fault environment, the filling of a reservoir drives the Mohr circle away from failure, thereby introducing stability. Jacob et al. (1976), in a (preprint) entitled "Tarbela Reservoir, Pakistan: A region of compressional tectonics with reduced seismicity upon initial filling" and the California data, seem to support Snow's theory.

Available evidence has tended to show a tensional or low stress environment for reservoir associated earthquakes in the Southeast United States. The tensional environment facilitates penetration of reservoir fluid into the rocks and lubrication and extension of existing structures. Data analyses of other reservoir areas are being used to determine if this generality can be applied to other areas. The hypothesis to be tested is that regions under a tensional stress state generally have ω^{-3} high frequency slope decay and are more susceptible to induced seismicity than areas under general compression.

V. Propagation Effects and Recording Limitations

Ideally, seismic data used for studying the high frequency spectral content of earthquakes should be recorded on instruments having a wide dynamic range and a broad-band frequency response. Such instruments have been used (Tucker and Brune, 1972), but they are expensive and difficult to operate in large numbers. The spectral study of Marion (1977) utilized a specially designed high-frequency (15 to 300 Hz) system to measure the slope in the 30 to 200 Hz range for very close (0.5 to 5.0 km) events.

Telemetry systems which are now used routinely for studies of small earthquakes have a typical dynamic range of about 40 db, two orders of magnitude, and a reasonably flat response to particle velocity for frequencies between 1 and 25 or 30 Hz (O'Neill and Healy, 1973). If the gains on a network are set to detect earthquakes of magnitude 0, the signals from earthquakes of magnitude 2 or greater can saturate the system. Also, the limited high-frequency response can tend to obscure the corner frequencies of the lower magnitude earthquakes that will be recorded within the dynamic range of the system.

The use of WWSSN short period data further reduces the range of usable data for spectra computation as the frequency response is narrow-band and the recorder turns at such a slow speed that it sometimes is almost impossible to follow the seismic trace.

Aside from recording characteristics, other effects can also degrade the seismic data and earthquake spectra. Perhaps most influences on the high frequency slope occur as a result of the propagation path taken. Surface wave contamination (Thatcher and Hanks, 1973), wave scattering (Dahlen, 1974), lateral inhomogeneities (Peppin, 1976) and most importantly, anelastic altenuation, are path effects which can bias the recorded data. Also, near-source effects (Peppin, 1976), multiple arrivals (Bakun and Bufe, 1975), and the selection of the seismic window studied (Murphy and LaHoud, 1975) can affect the frequencies seen.

The method used in digitizing the seismic trace can also affect the spectra. Noise may dominate at high frequenices if too wide a digitizing interval is used; or, at low frequencies, if these frequencies are absent from the digitized portion. Frequencies, with sufficient amplitude, higher than the Nyquist (or folding) frequency can introduce lower frequencies into the spectra.

P and S-wave spectra are commonly used in the determination of seismic source parameters. For several reasons, we prefer to limit our analysis to P-waves. The P-wave is attenuated less than the S-wave and travels faster. The P-wave amplitudes are typically lower than amplitudes of S-waves and hence P-waves are less likely to be saturated. Presumably, the earliest arrivals more accurately reflect the source while the slower, later arrivals reflect near-surface structure (Peppin, 1976). This is supported by Hanks and Wyss (1972) who found P-waves to be preferable to S-waves for spectral analysis.

In an exercise to identify the range of valid earthquake data, characteristics of the recording system, along with expected amplitudes and corner frequencies for certain magnitude events, will be combined with the effects of attenuation. To derive the spectral slope beyond the corner frequency, frequencies recorded to about 4 times the corner frequency are needed. In addition, to preserve the source effects and keep the signal to noise ratio high, amplitudes which are greater than or equal to 0.2 of the original amplitude are desired. Utilizing the mentioned factors, the size and range of events that can be successfully analyzed for high-frequency slope are determined.

For the study of the high-frequency slope to be useful in interpreting source conditions, it is necessary to allow for some of the factors that can alter the data. The single largest problem in analyzing the spectra of body waves for source characteristics is accounting properly for path effects that change the earth's frequency response between source and station. The shape of high-frequency spectra is severely affected by attentuation varying with path depth. The form for seismic attenuation ϕ is usually given as

$\phi(f,x) = \exp -(\pi fx/Qv)$ where f = frequency

x = distance in km, Q = specific attenuation factor

v = wave velocity, km/sec

We assume all anelastic attenuation effects can be accounted for by a frequency independent Q. That Q is found to be frequency dependent by some investigators (Sumner, 1967; Jackson and Anderson, 1970; Solomon, 1972; Aki and Chouet, 1975) is probably not crucial for the local earthquakes used in this study. The value of Q for the crystalline rocks of the earth's crust is usually found to be 100 or greater (Long and Berg, 1969; Clowes and Kanasevich, 1970; Hill, 1971). Accordingly, in figure 2 we treat P-wave Q values of 200, 500, and 800 (to encompass various attenuation levels) at assumed distance to obtain the maximum observable frequency consistent with the condition that $\exp-(\pi ft/Qv)\geq 0.2$. The higher frequencies attenuate rapidly for all values of Q. This is in general agreement with the finding of Tanis (1973) that no reliable seismic information higher than about 30 Hz is found in the range of 10 to 100 km.

As evaluation of the spectral slope past the corner frequency is the desired end product of this analysis, we will consider relations between corner frequency and magnitude. Bakun et al. (1978), in their study of the high-frequency radiation of small earthquakes, implied that corner frequencies vary systematically with azimuth and

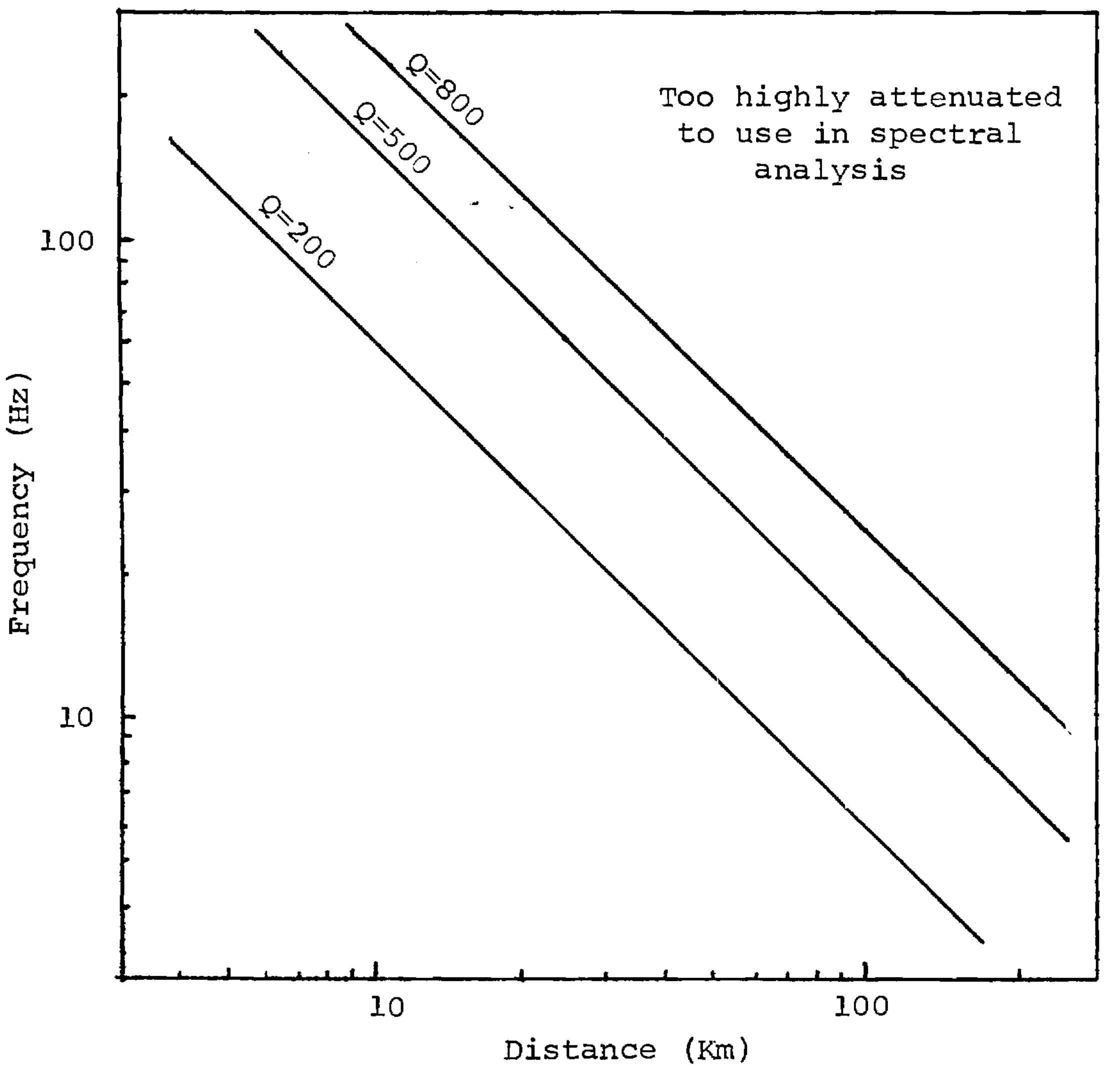

Figure 2. The effect of Q on the frequency of seismic waves with distance using the attenuation formula $\exp -(\pi fx/Qv) \geqslant 0.2$ of the original energy. Frequencies above each line are too highly attenuated to use reliably in spectra computation.

takeoff angle. Figure 3 shows an average corner frequency versus magnitude relation. However, it should be noted that data spread is often wide. As the spectral slope out to about 4 times the corner frequency is needed to estimate the decay, computation of slopes of events with corner frequencies above average will be close to or beyond the capabilities of the recording instruments.

In figure 4, the results of figure 2 and figure 3 are combined to obtain a minimum magnitude with distance for which attenuation is sufficiently slight to allow determination of the spectral slope. An equivalent empirical formula was obtained by Bakun and Bufe (1975). They found for southern California earthquakes that if the Q value was as low as 4(fĉ)t where fc is the corner frequency and t the travel time, the corner frequency is determined by path effects.

The gain of the seismic system and the effects on the data must also be recognized. If the gains of a network are set high, the signals of the larger earthquakes will clip in the recording system. Typical telemetry systems use a gain of 100 K or 200 K and World Wide Standard Seismograph Network short period stations commonly use 50 K magnification. Examination of seismic records of these two systems indicates a maximum unsaturated amplitude of 40 mm (zero to peak) for telemetry systems and 20 mm (zero to peak) for WWSSN short period data.

To recognize the maximum magnitude which can be used without saturation of the records, we calculate the expected amplitude with distance for 200 K, 100 K and 50 K magnifications. Richter's magnitude $M_L = \log (A) - \log (A_o)$ (1958) for southern California earthquakes and Nuttli's magnitude $m_b = 3.75 + 0.9 \log 10 + \log$

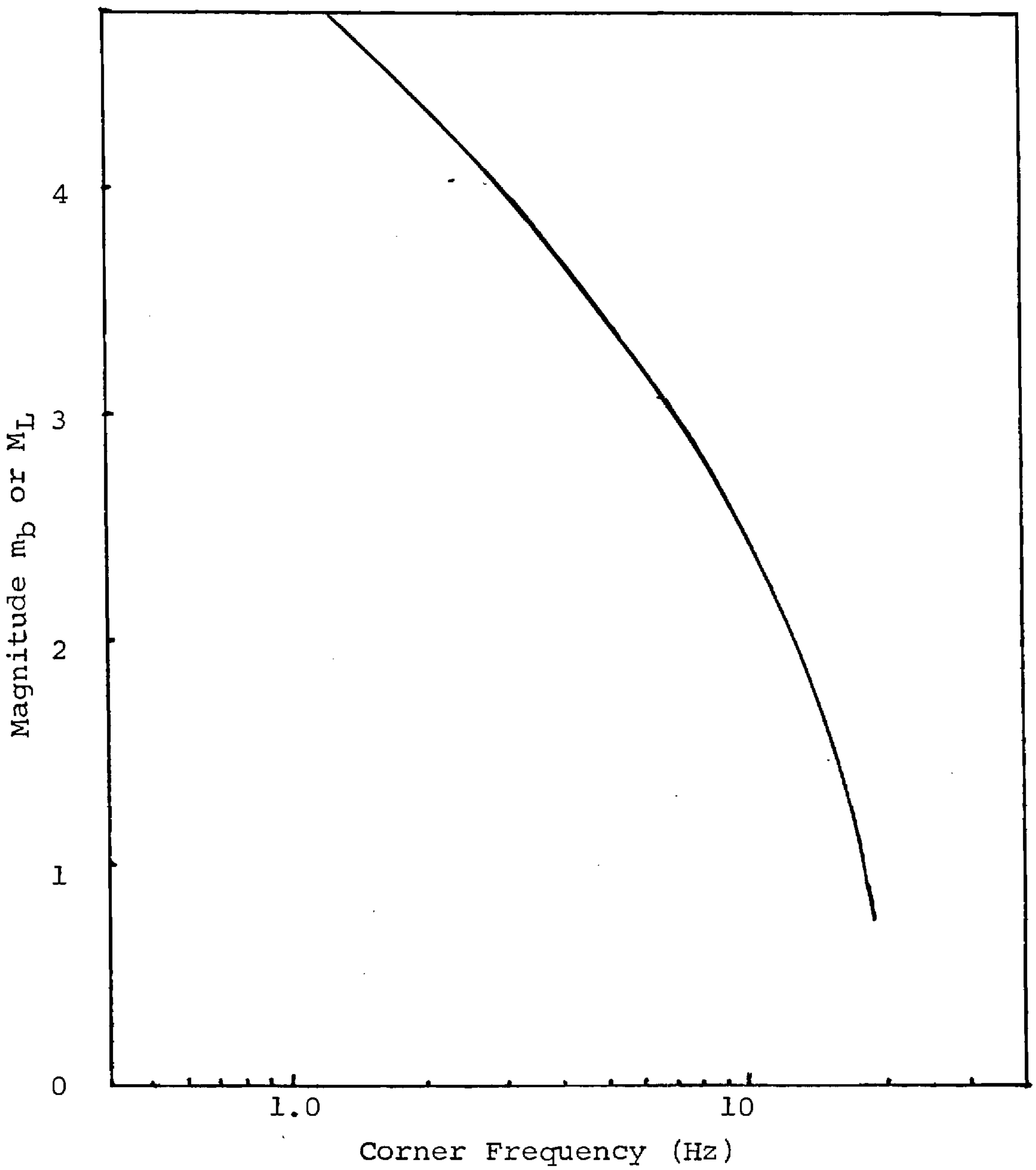

Figure 3. Corner frequency f_c versus magnitude m_b or M_L from Frasier and North (1978), O'Neill and Healy (1973), Tanis (1973) and Aki and Chouet (1975).

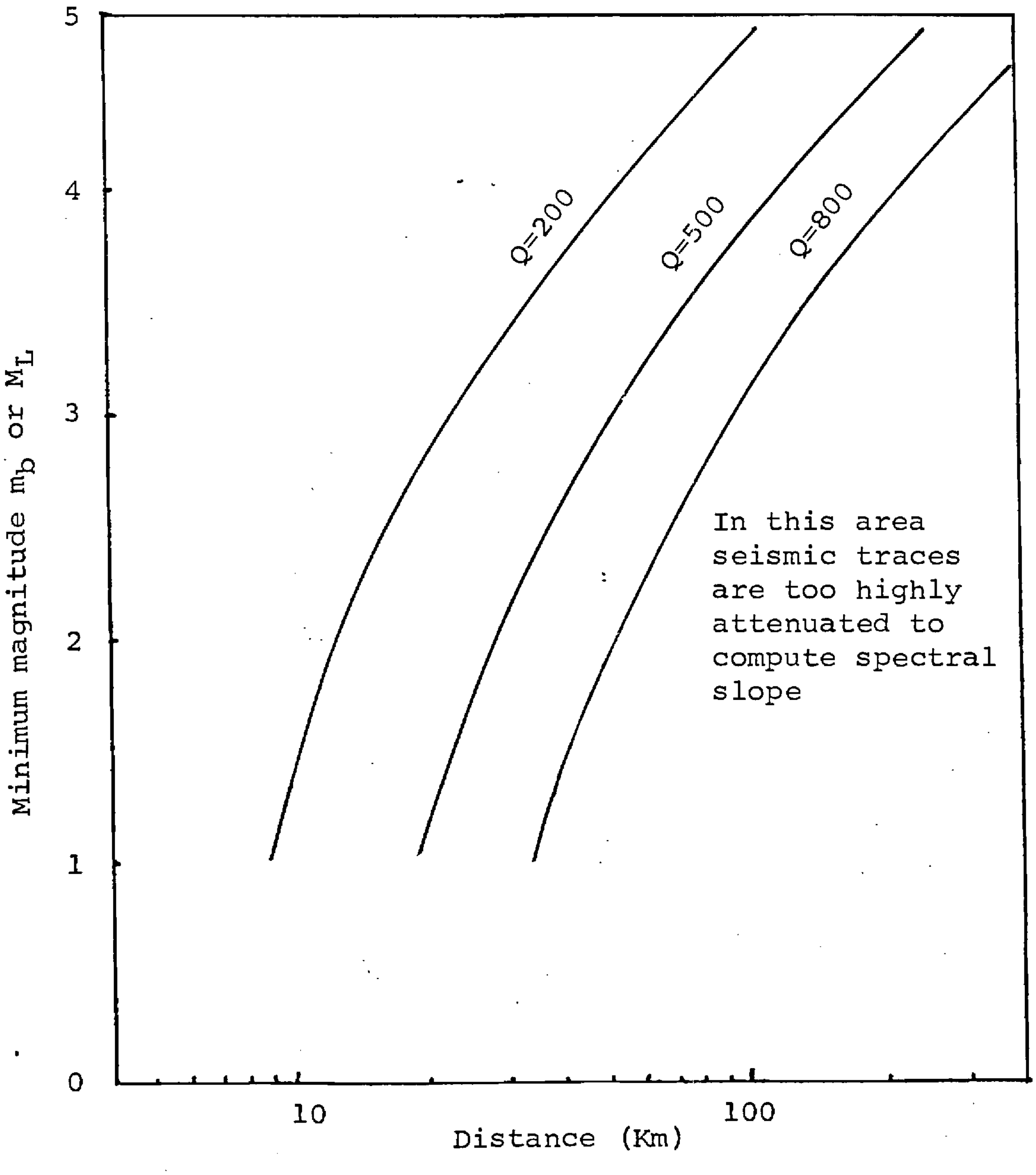

Figure 4. Minimum magnitude with distance for different Q values that can be used to evaluate the spectral slope at 4 x f_c without attenuation effects.

$_{10}$ (A/T), $0.5^{o} \leq \Delta \leq 4^{o}$ (1973) for Eastern North America are widely used. Figures 5 and 6 show expected amplitudes with distance for the Richter and Nuttli magnitudes.

The knowledge of the maximum amplitude that can be recorded without saturation can be combined with the results represented by figures 5 and 6 to obtain a maximum magnitude versus distance that can be recorded successfully without loss of the signal. Figures 7 and 8 represent the maximum magnitude versus distance that can be recorded at 50 K, 100 K and 200 K for Nuttli's m_b and Richter's M_L magnitudes respectively.

The instrument response for typical telemetry systems and World Wide Standard Seismograph Network short period records are now considered. Figure 9 shows the frequency versus displacement response of the Georgia Institute of Technology telemetry system at 200 K and the WWSSN short period at 50 K magnification. The telemetry system records frequencies primarily from 4-45 Hz at 0.3 of the maximum velocity response. The WWSSN SP records primarily 0.2 - 5.0 Hz at 0.1 of the maximum velocity response.

Combining all the previous analyses, the range of earthquakes for which the high frequency slopes past 4 times the corner frequency can be successfully recorded can be determined. In figures 10-13 the usable range of earthquakes as determined by saturation, anelastic attenuation, and the frequency response of the seismic system used are shown. Similar analyses could be done for other recording systems or for different attenuation levels.

As noted, recording systems having a wide dynamic range and broadband frequency response such as used by Marion (1977) to

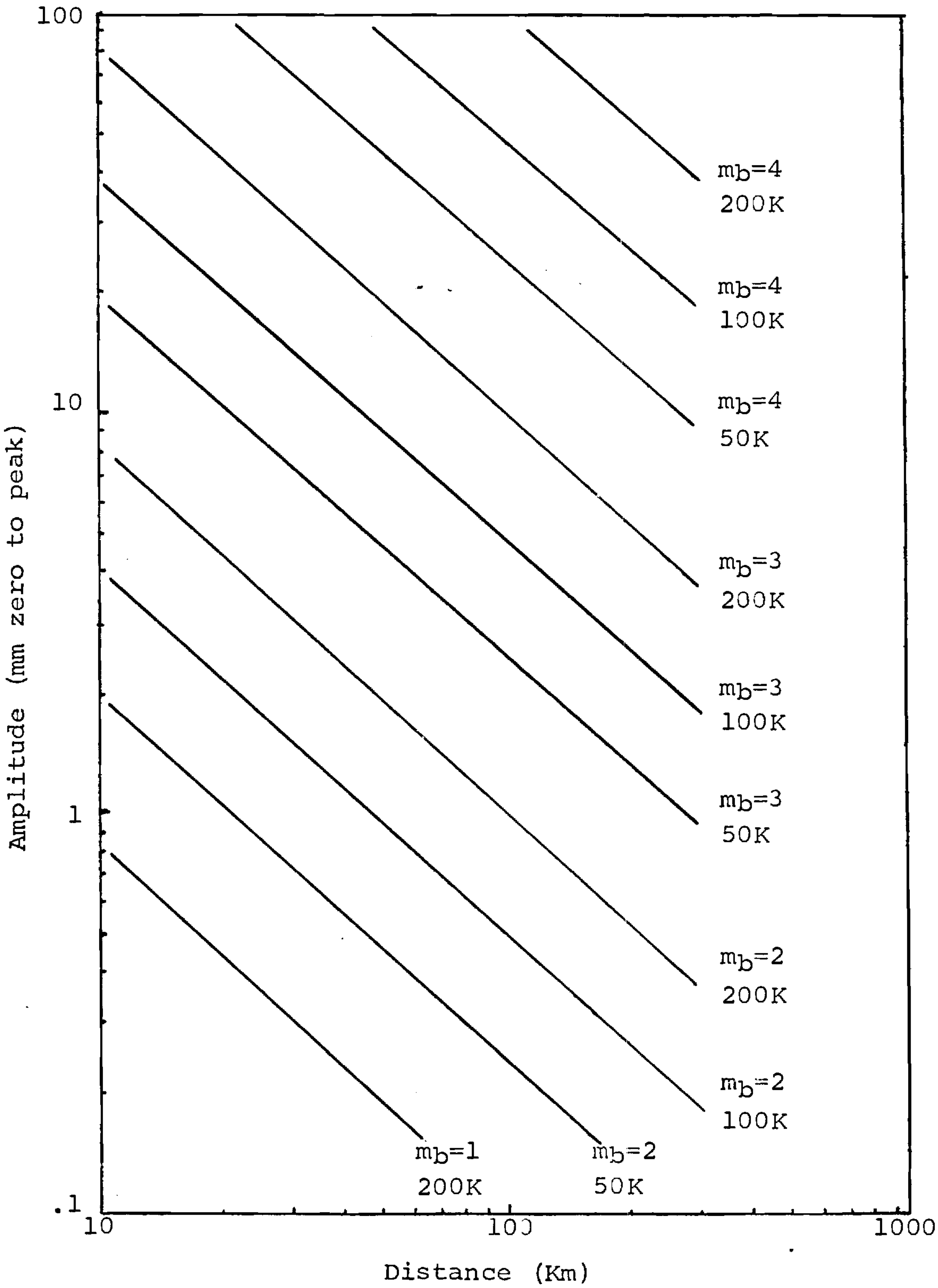

Figure 5. Expected sustained amplitude with distance for Nuttli's m_b (1973) for P-waves $\leq$ 1/4 Lg phase amplitude.

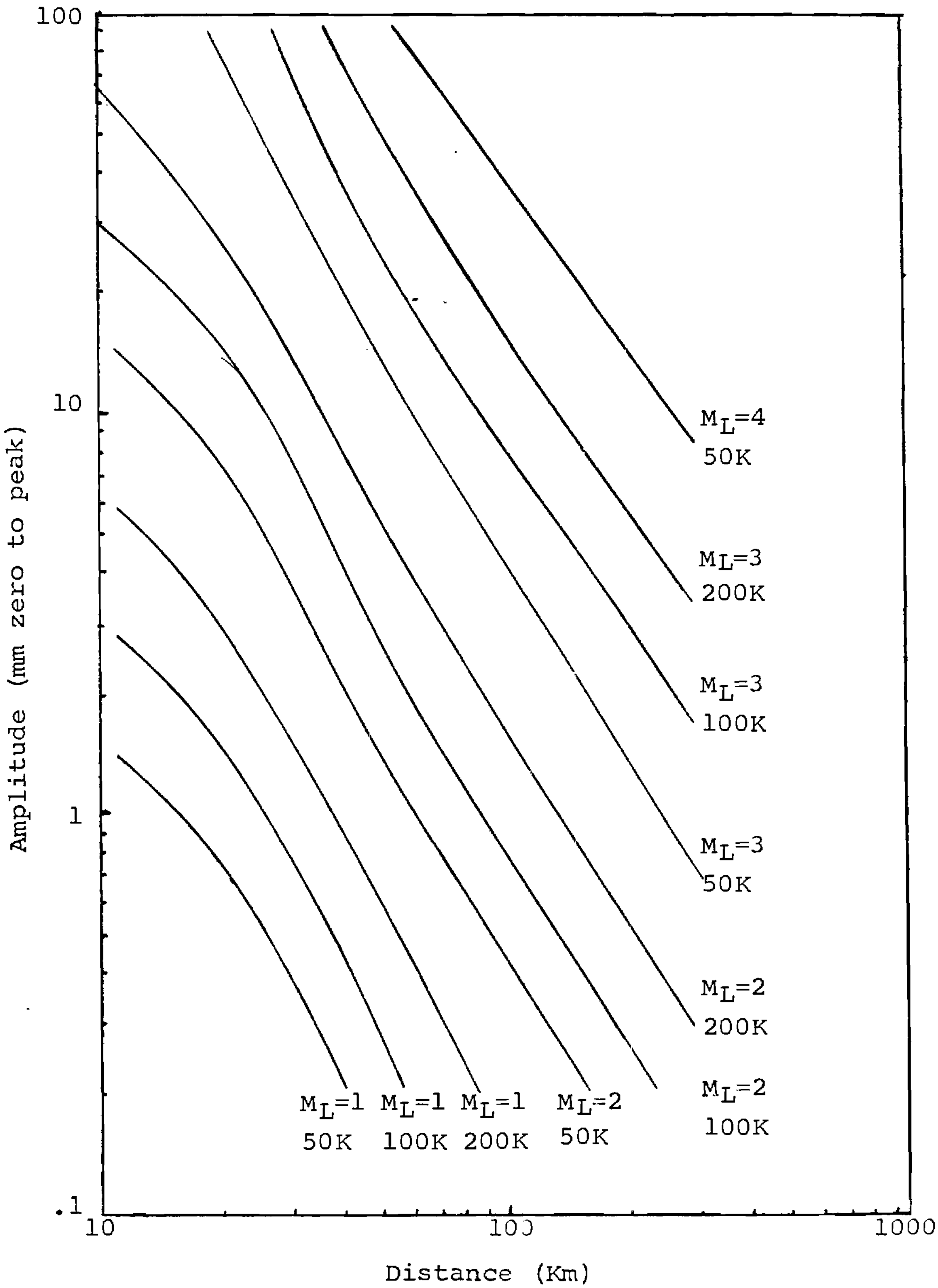

Figure 6. Expected amplitude with distance for Richter's M_L (1958) for P-waves $\leq$ 1/4 surface wave amplitude

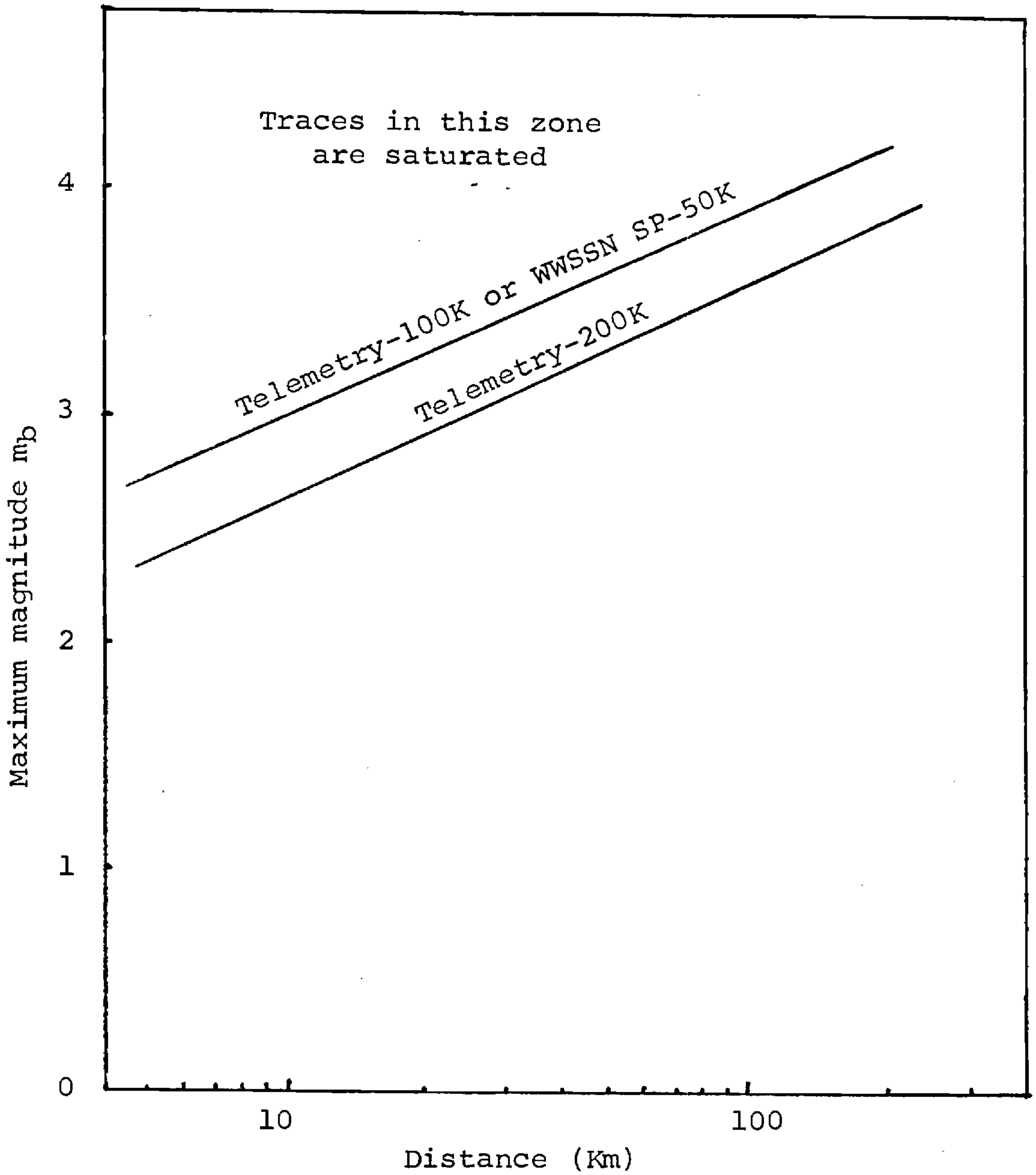

Figure 7. Maximum magnitude m_b versus distance to avoid saturation of the seismic signal. Regions above each line are too saturated for spectral computation.

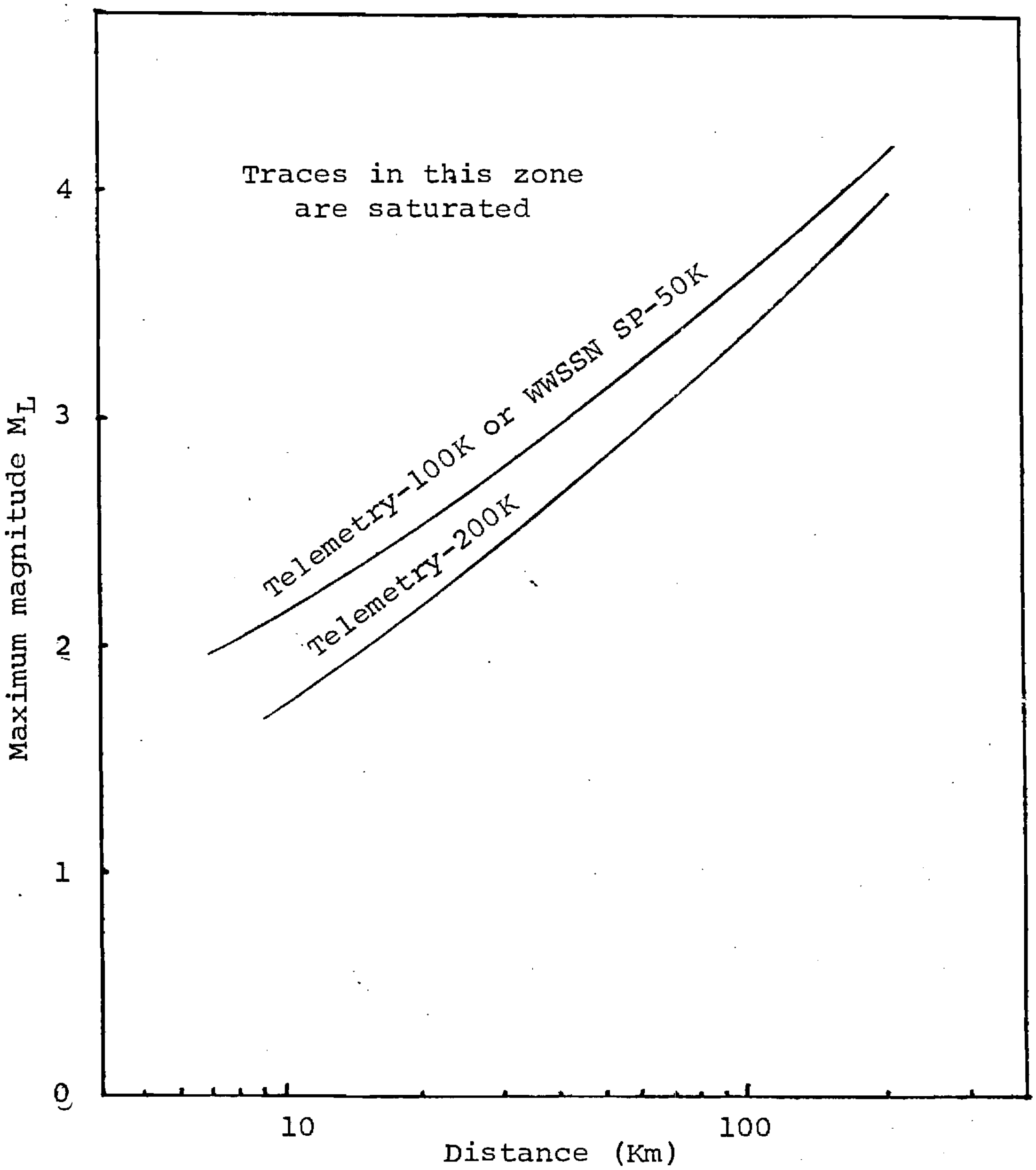

Figure 8. Maximum magnitude M_L versus distance to avoid saturation of the seismic signal. Regions above each line are too saturated for spectral computation.

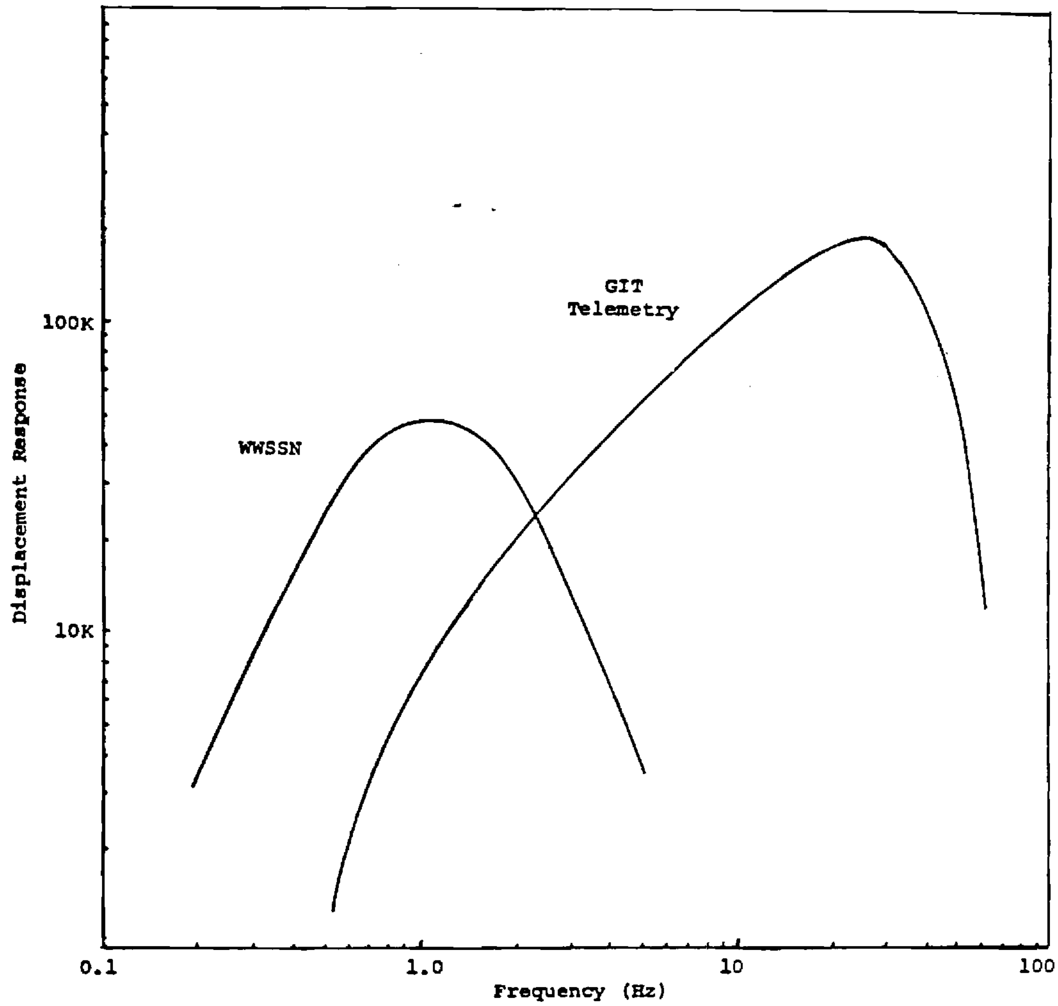

Figure 9. Frequency response of seismographs.

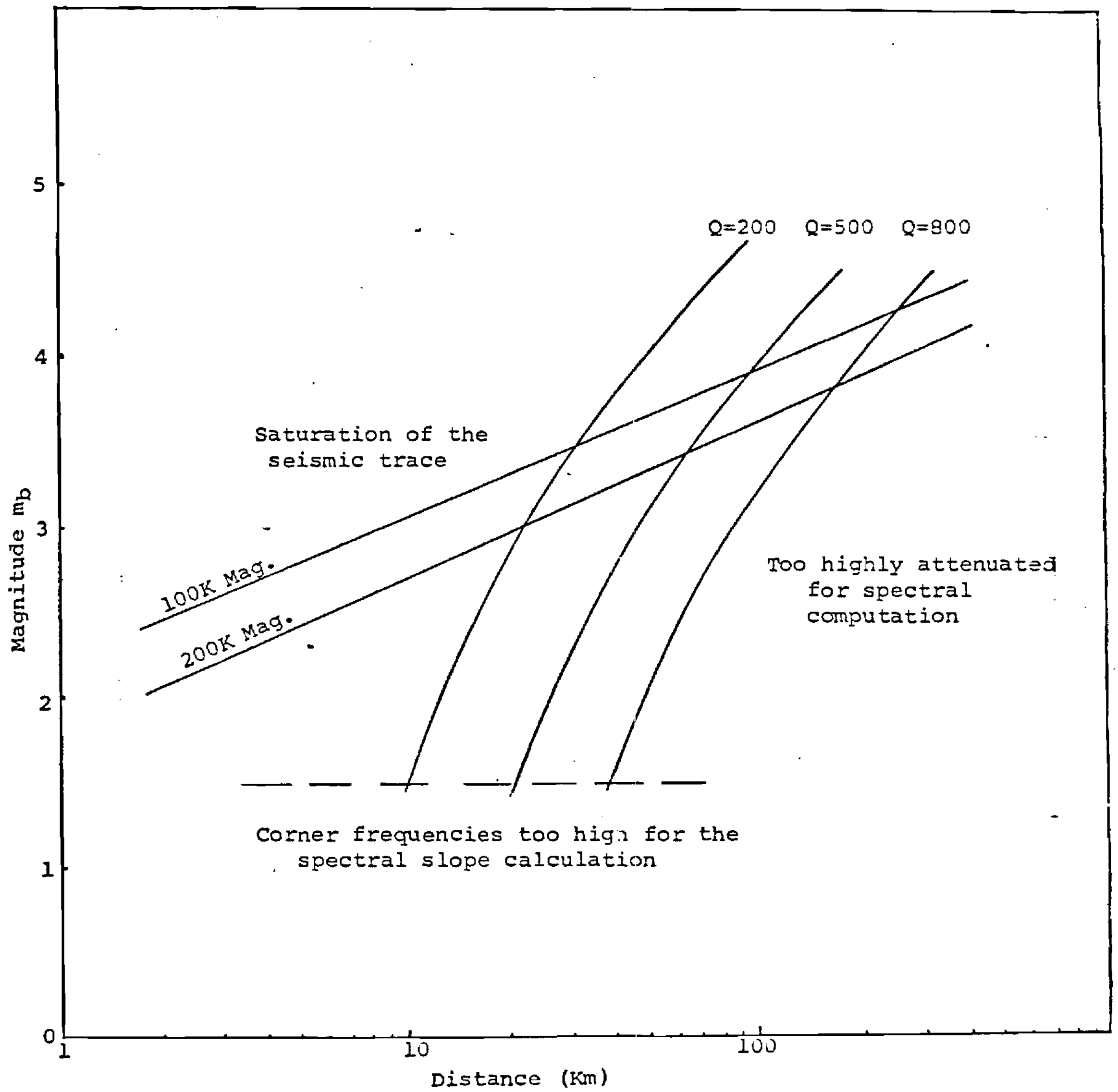

Figure 10. Range of magnitudes versus distance that can be used for determination of the spectral slope past the corner frequency for a typical telemetry system using Nuttli's m_b.

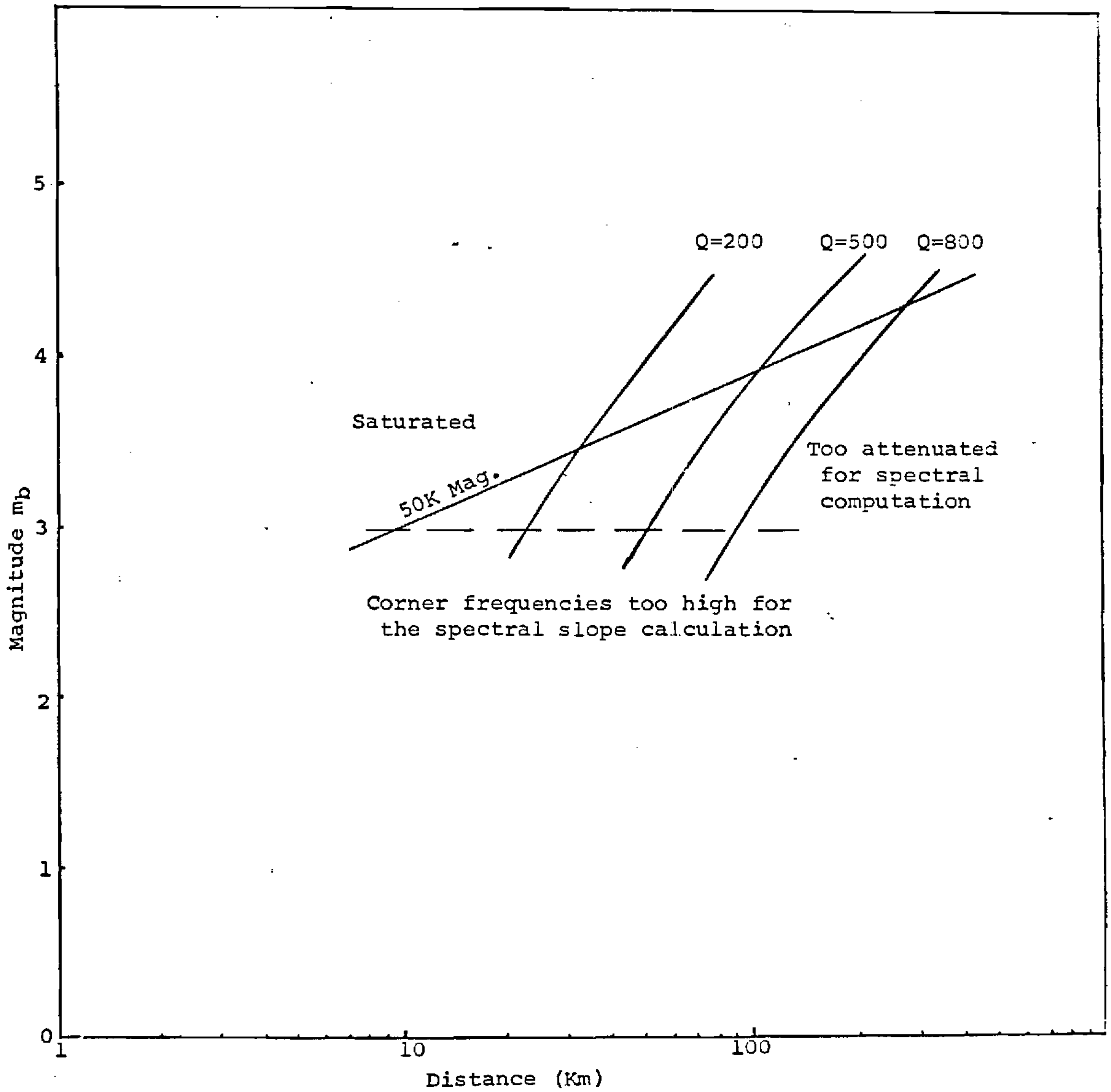

Figure 11. Range of magnitudes versus distance that can be used for determination of the spectral slope past the corner frequency for WWSSN short period using Nuttli's m_b.

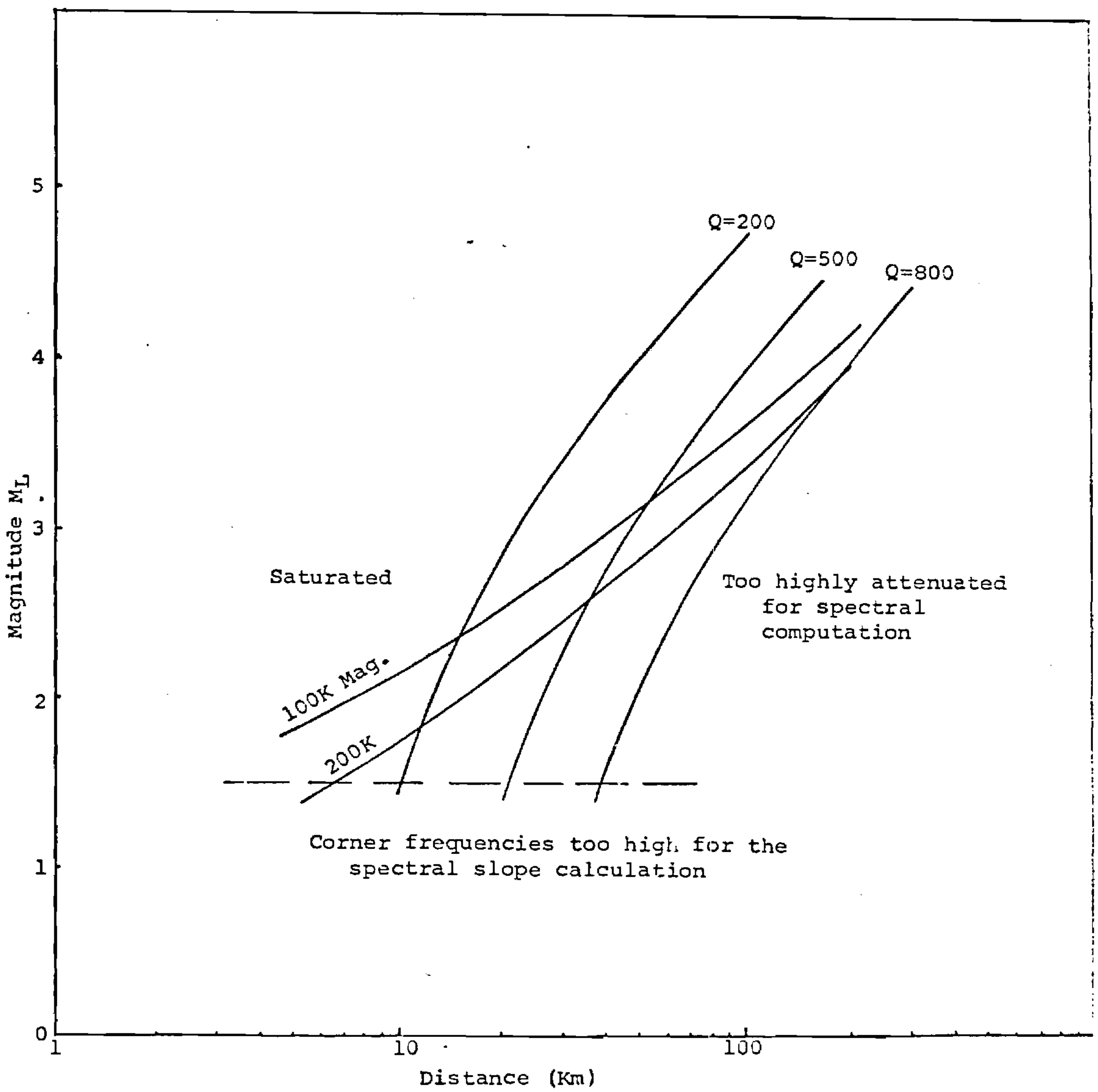

Figure 12. Range of magnitudes versus distance that can be used for determination of the spectral slope past the f_c for a typical telemetry system using Richter's M_L.

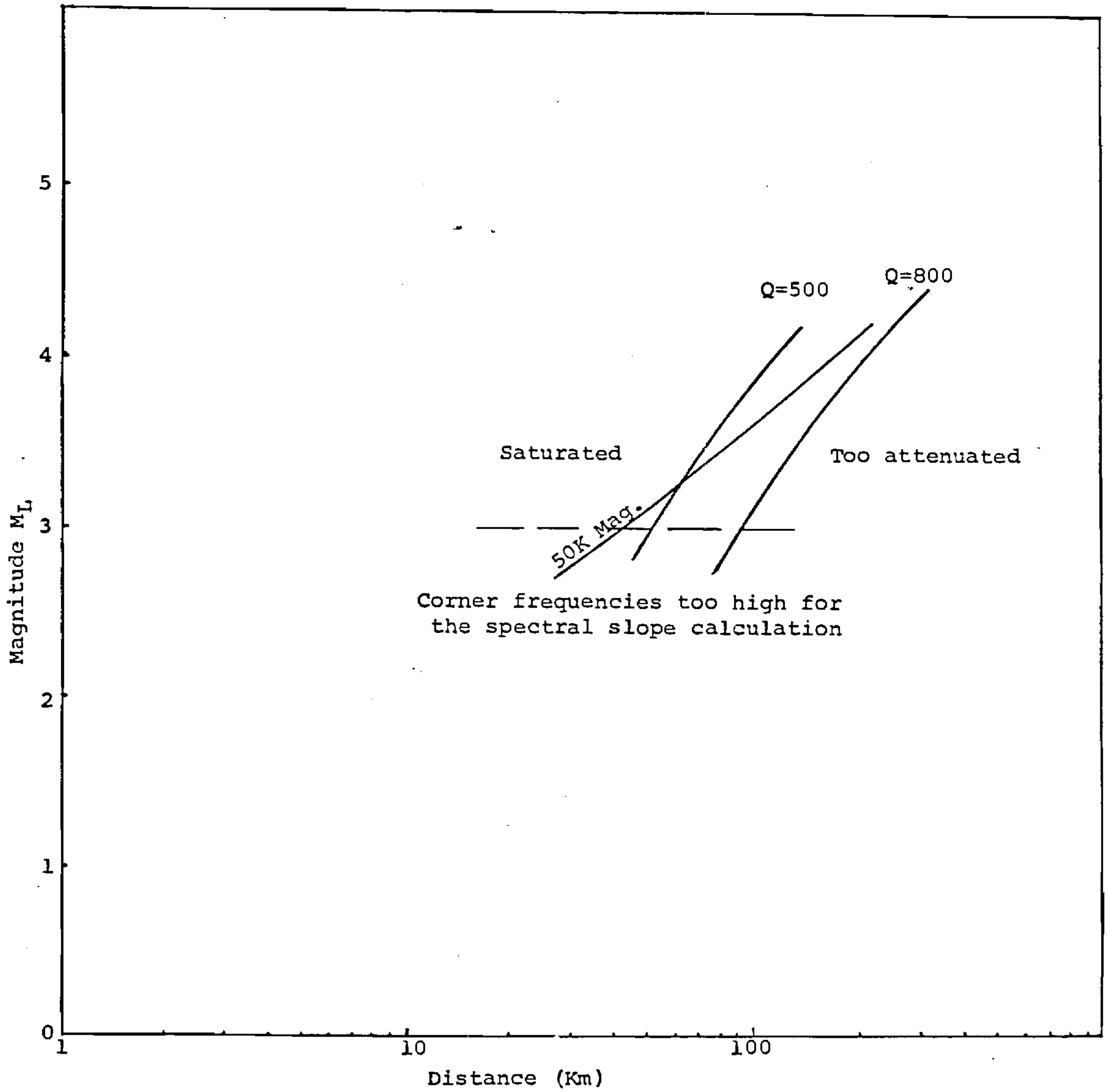

Figure 13. Range of magnitudes versus distance that can be used for determination of the spectral slope past the f_c for WWSSN SP using Richter's M_L.

record very close events are ideal for high-frequency spectral study. As recordings by systems such as these are not common or readily available, the study will primarily depend on obtaining telemetry system records which fit the derived criteria.

VI. Data Regions

For this study to be significant, data from reservoir areas that have exhibited induced seismicity must be compared to data from reservoir areas with naturally occurring events. A problem is that naturally occurring earthquakes near reservoirs which have not experienced induced seismicity are often not reported. Also, as noted, most reservoirs are not instrumented to record local earthquakes. These factors limit the number of possible data areas. However, on a worldwide basis, the number of reservoirs reporting seismicity is significant although some of these seismic records do not lend themselves readily to spectral analysis. Also, the large number of instrumented reservoirs in the United States could be a vital factor in obtaining suitable data.

In Table 2, possible data regions for obtaining copies of reservoir associated earthquakes are listed. While this table hopes to encompass all reported seismic activity in reservoir areas, obviously new reports or reports unknown to this study are anticipated.

The results of Table 2 indicate suitable data for spectral analysis of reservoir associated earthquakes must of necessity came from the United States and possibly Canada. Data from reservoir areas in the Southeastern United States, California, and Lake Mead are expected to furnish the majority of the data used in this

Table 2. Possible Data Regions

United States

	Reservoir	Activity	References	Data Availability
1.	Hoover Dam (Lake Mead) Nev. - Az.	1938, Mag 5 1938-present micro eq's.	Carder (1945) Carder (1970) Hofmann (1973) Anderson and Laney (1975)	Telemetry network
			Lee and Matamoros (1975)	Wrote A. M. Rogers requesting 5-10 seismic records
			Rogers and Lee (1976)	
2.	Oroville Dam California	1975, Mag 5.9	Hofmann (1973) Bufe et al. (1976)	U.S.G.S. Branch of Ground Motion and Faulting for high gain data
			Morrison et al. (1976)	Contacted Dr. Bufe for data Wrote Mr. Baker
3.	Clark Hill GA.-S.C.	1974, Mag 4.3 1974-present, Micro eq's.	Talwani (1976)	Marion (1977) GIT tapedecks GIT telemetry
4.	Jocasse S. Carolina	1975, Mag 3.2 1975-present micro eq's.	Talwani (1978)	Marion (1977) GIT tapedecks USGS or USC telemetry
5.	Keowee S. Carolina	1971, MM-V 1971-present micro eq's.	Talwani (1978)	GIT tapedecks USGS or USC telemetry
6.	Monticello S. Carolina	1977-present, micro eq's.	Talwani (1978)	GIT tapedecks USGS or USC telemetry
7.	Long Valley California	micro eq's. not associated with reservoir	Daly et al. (1977)	Dr. Malcolm Clark USGS, Menlo Park, CA Dr. Rob Wesson USGS, Reston, VA
8.	Sanford Dam Texas	micro eq's.	Shurbet (1969)	not available, analog
9.	Cabin Creek Colorado	micro eq's.	Evans (1966)	not available
10.	Chatfield and Bear Creek Colorado	perhaps post impoundment	Major in Patrick (1977)	Contact Dr. Major at Colorado School of Mines

	Reservoir	Activity	References	Data Availability
11.	Palisades Idaho	micro eq's	Sbar *et al*. (1972) Schleider (1975)	Smoked paper
12.	Kerr Dam Montana	4/69-4/71 54 events Mag 2.0-4.8	Dunphy (1972)	station characteristics and availability not known
13.	San Luis Dam California	micro eq's not associated with reservoir	Hofmann (1973) Mickey (1973)	Same as Long Valley Dam
14.	Flaming Gorge Utah Glen Canyon Arizona	1960's, some natural activity	Mickey (1973)	Check with U.S.G.S.
15.	Cedar Springs California	1960's, some natural activity	Mickey (1973)	Same as Long Valley Dam

Algeria

Oued Fodda	1933, micro eq.	Gupta and Rustogi (1976)	not available

Australia

Talbingo	1971, mag 3.5	Muirhead et al. (1973) Timmel and Simpson (1973)	3 portable stations installed 7/71; write Muirhead or Timmel
Eucumbene	1970, mag 5.0	Timmel and Simpson (1973) Gupta and Rastogi (1976)	Write Timmel

Austria

Schlegeis	1971, micro eq	Blum and Fuchs (1974)	No, eq's $\leq$ mag 1.0

Canada

Mica McNaughton Lake, B.C.	1973, Swarm with mag = 4.7	Ellis *et al*. (1976)	Wrote to Dr. Ellis no response

Canada (continued..)

Reservoir	Activity	References	Data Availability
Manic 3	1975, microeq's Largest =4.3	Milne and Berry (1976b) LeBlanc and Anglin (1978)	Contacted Gab Leblanc Wrote to Frank Anglin for data

China

Reservoir	Activity	References	Data Availability
Hsinfengkiang	1962, mag 6.1	Wang *et al.* (1976)	Probably not

Ethiopia

Reservoir	Activity	References	Data Availability
Tendaho	instrumented at press time	Lane (1974) Long (1974)	(?) on magnetic tape

France

Reservoir	Activity	References	Data Availability
Monteynard	4/25/63 Mag 4.9	Gupta and Rastogi (1976) Rothe (1968)	Seismic station at Roseland, 110 km away too far
Grandval	8/5/63 Intensity V	Gupta and Rastogi (1976)	Write Rothe (?)
Vouglans	6/21/71 mag 4.5 followed 20 aftershocks	Bozovic (1974)	Write Rothe (?)

Greece

Reservoir	Activity	References	Data Availability
Kremasta	1966, Mag 6.3 1/15-1/20/66 6 to 8 tremors	Gupta and Rastogi (1976)	Write Papazachos for records of Valsamata, 90 km away
Marathon	1938, 2 eq Mag 5.0	Galanopoulos (1967) Gupta and Rastogi (1976)	Not available

India

Reservoir	Activity	References	Data Availability
Koyna	1967, mag 6.5	Gupta and Rastogi (1976)	Ordered 31 seismograms of mag 3.5-4.2,Station POO, 115 km
Several in Southern India	Microeq's	Guha *et al.* (1974) Gupta and Rastogi (1976)	Write Guha

Iran

Reservoir	Activity	References	Data Availability
Lar	Instrumented at press time	Lane (1974) Long (1974)	(?) on magnetic tape

Italy

Reservoir	Activity	References	Data Availability
Vajont	early 1960's microeq's	Gupta and Rastogi (1976) Caloi (1966)	No
Pieve Di Cadore	1950, micro	Bozovic (1974)	No
Piastri	4/7/66-10/66 Intensity IV-VII	Bozovic (1974)	Detailed data lacking

Japan

Reservoir	Activity	References	Data Availability
Kurobe	microeq's	Hagiwara and Ohtake (1972)	No, WWSSN type at close range
Kamafusa	1970, micro max = 2.5	Gupta and Rastogi (1976)	recording uncertain
Matsushiro injection	micro max = 2.8	Ohtake (1974)	magnetic tape

New Zealand

Reservoir	Activity	References	Data Availability
Benmore	1966, mag 5.0	Adams (1974)	Wrote to R.D. Adams, reply by Mr. Eiby was that data not available
Pukaki	Monitored but not known	Adams (1974)	Same as above

Pakistan

Reservoir	Activity	References	Data Availability
Mangla	1965-1968, microeq 10/70 - mag4.2	Brown (1974) Adams (1972)	recording sensitivity and freq. response not adequate
Tarbela	reduced	Simpson (1976) Jacob *et al*. (1976) Armbruster *et al*. (1978)	Write to K.H. Jacob

Rhodesia

Reservoir	Activity	References	Data Availability
Kariba	1963, mag 5.8	Gough and Gough (1970)	Wrote to Gough Wrote to J. N. Allen requested 10 records

South Africa

Reservoir	Activity	References	Data Availability
Hendrik Verwoerd	1970, micro max. = 2.0	Gupta and Rastogi (1976) Green (1974)	No, only regional network

Spain

Reservoir	Activity	References	Data Availability
Cannalles	1962, Intensity V	Gupta and Rastogi (1976)	No
Camarillas	12/61, max micro	Bozovic (1974)	No
El Grado	(?)	Yague (1969)	No

Switzerland

Reservoir	Activity	References	Data Availability
Contra	10/65-11/65 micro	Gupta and Rastogi (1976)	No

Turkey

Reservoir	Activity	References	Data Availability
Keban	1974, mag 3.5	N. Tilford (1975) in Simpson (1976)	unknown

Uganda

Reservoir	Activity	References	Data Availability
Murchison Falls	micro	Lane (1974)	No

U.S.S.R.

Reservoir	Activity	References	Data Availability
Nurek	1973, mag 4.0-4.5	Nikolaev (1974)	unknown

Yugoslavia

Reservoir	Activity	References	Data Availability
Bajna-Casta	7/3/67 mag 4.5	Simpson (1976) Bozovic (1974)	unknown, national network
Grancarevo	mag 1-2	Bozovic (1974)	unknown
Bileca	increased	Gupta and Rastogi (1976)	unknown

study. The proximity of Canada along with some reports of induced seismic activity there could be beneficial although at this time data availability is not known.

VII. Proposed Additional Tasks

During the second half of the project we expect to use data we have requested or possibly obtained through field measurements to compute additional spectral slopes. Because of the wide variations in Q that can exist and because of the strong dependence of the spectra under some conditions on Q, we expect to use our data to compute Q on a regional basis. During analysis we intend to document the traditional method for spectra computation and test modifications utilizing averaged estimates or maximum entropy methods for application when the data prohibit the traditional analysis. Finally, we hope to evaluate the potential usefulness of the -cube discriminant and make recommendations for its utility.

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Final Technical Report

A SEISMIC SPECTRAL DISCRIMINANT FOR RESERVOIR INDUCED EARTHQUAKES

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Final Technical Report

"A Seismic Spectral Discriminant for Reservoir Induced Earthquakes"

Leland Timothy Long and Greg Johnston

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Project Summary

The object of this research has been to evaluate a spectral discriminant which might allow identification of other areas where new reservoirs would induce significant seismic activity. The discriminant is the high-frequency slope of the displacement spectra of earthquakes occurring in the vicinity of a new or proposed reservoir. An ω-cube slope from earthquakes in the vicinity would predict that nearby reservoirs would induce earthquakes and an ω- square slope from earthquakes in a region would predict that reservoirs in the vicinity would not induce earthquakes. This association of ω-cubic decay with induced reservoir seismic activity and ω-square decay with areas where reservoirs do not induce seismic activity was observed first in studies of seismic spectra of Southeastern United States earthquakes. Earthquakes occurring in the Folded Appalachians have an ω-square high-frequency decay and reservoirs in that region are not known to induce seismic activity. Earthquakes observed in two Piedmont Province reservoirs which have induced seismic activity show ω-cubic spectral decay. Hence, the limited data for the Southeastern United States indicated that the high-frequency spectral decay may be a viable discriminant for the identification of other areas susceptable to induced seismic activity. In this study one objective was to test the generality of the discriminant for identifying other areas of induced seismic activity.

Fundamental to this study was an investigation of whether there exists a theoretical basis for the discriminant. Most theoretical studies on the radiation of high-frequency seismic energy indicate that the high-frequency energy is controlled to a large extent by the slip velocity on the fault. At the rupture front, velocity changes which may be related to variable stress conditions or fault irregularities enrich the spectra with high-frequency energy. In contrast, a lubricated fault plane moving under low-compressive stress conditions would not possess the same intensity of high-frequency energy. Hence models of the former predict ω- square decay and models of the latter predict ω-cube decay. The latter corresponds to conditions expected for the shallow reservoir-induced earthquakes.

In our literature review only two reliable examples of spectra were found from which we could measure the high-frequency slope. Detailed spectra from the Oroville aftershock sequence show ω-cube decay giving credence to the claim that these events were reservoir induced. Spectral studies of Lake Mead earthquakes showed S-wave high-frequency decay of $\omega^{-1.2}$ to $\omega^{-1.8}$ but the conclusion of the authors was that these were natural, rather than induced, events. In an attempt to supplement the data we obtained new spectral data from southeastern United States reservoir areas. Seventeen new spectra from the Monticello (S.C.) reservoir area and 32 spectra from the Clark Hill (Georgia-South Carolina) Reservoir area were obtained. These spectra showed $\omega^{-3.0}$ decay and are from areas of induced reservoir activity. We also obtained 22 new spectra from the Lake Sinclair, Georgia, area, an area of suspected but unproven induced seismic activity. These spectra show generally a $\omega^{-2.0}$ decay. The Wallace Dam (Lake Ocoee) on the eastern branch of Lake Sinclair has been carefully monitored for seismic activity but none has been detected after one year of loading.

We were unable to come to definite conclusions concerning the spectral discriminant because of the sparsity of data. Where appropriate data could be examined, the spectral slope did discriminate areas where reservoirs induce earthquakes from areas where reservoirs do not induce earthquakes. Theoretical models of the seismic source allow a rational explanation for the discriminant based on the character of the rupture velocity and the relation between frictional resistance and driving shear stress. Models which are based on a uniform, transonic rupture along a smooth circular fault best satisfy the spectra from areas of reservoir induced seismicity. We uncovered no contradictory evidence that could not be explained. The success of the limited data and the theoretical rational are compelling circumstantial evidence supporting the discriminant.

We recommend two to three years of high-quality digital recording of seismic data prior to the filling of reservoirs. We recommend also the evaluation of spectral signatures as a function of depth to test whether this might be a significant factor in the apparent success of the discriminant.

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A SEISMIC SPECTRAL DISCRIMINANT FOR RESERVOIR INDUCED EARTHQUAKES

I. Introduction

The correlation between the creation of some water storage reservoirs and increased seismic activity has been widely recognized in recent years. Earthquakes which have caused damage to structures and loss of life have occurred near large reservoirs at Kariba in Rhodesia, at Kremasta in Greece, and at Koyna in India. At this time, 66 cases of reservoir associated seismicity have been reported worldwide (Parker *et al*., 1978; Talwani *et al*., 1978; Buforn, 1979). The largest earthquakes and longest duration of induced seismic activity has occurred at the deep or very large reservoirs.

An increase in the number of reservoirs with induced seismicity is expected as over 50 additional deep or very large reservoirs are to be completed by 1985 (Packer *et al*., 1978). Earthquakes with the potential to cause damage could occur in some of these new reservoir areas or in previously aseismic reservoirs. If reservoir associated earthquakes should occur in areas of high population density, where a previously low historical seismicity has left the population and structures unprepared for a damaging earthquake, the potential for damage could be great.

Increases in seismic activity near some reservoirs in the Piedmont Province of the southeastern United States has been reported in recent years. Earthquakes near Lake Jocassee, Lake Keowee and Monticello Reservoir in South Carolina have been attributed to reservoir impounding (Talwani, 1978). The Lake Jocassee area where seismicity was believed to be decreasing for nearly 3 years, recorded its largest earthquake,

m_b = 3.6, on August 25, 1979. The Clark Hill Reservoir area on the Georgia - South Carolina border, may also be a site of induced reservoir seismicity. Sporadic activity (m_b = 1.8- 3.0) from the Clark Hill Reservoir area has been detected at WWSSN ATL since 1963. On August 2, 1974 the largest recent event, a magnitude 4.3 earthquake, occurred in the northern portion of the Clark Hill reservoir.

Many more reservoirs in the southeastern United States have caused no appreciable or detectable seismic activity. Two of these reservoirs (Lake Ocoee and Carters Lake, Georgia) were monitored specifically for local induced earthquakes but, no unusual seismic activity has occurred. In a study of southeastern United States earthquakes, Marion (1977) found that the high frequency displacement spectra of earthquakes located near the Clark Hill and Jocassee reservoirs in the Piedmont Province had a different character than spectra of earthquakes in the Folded Appalachian area where no association between earthquakes and reservoirs is observed. A high-frequency displacement slope of ω^{-3} was measured in earthquakes which occurred near Clark Hill and Jocassee reservoirs in the Piedmont Province. Earthquakes of the Folded Appalachian Province were found to have an ω^{-2} high-frequency displacement decay. The ω^{-3} high-frequency spectral slope can be physically related to the earthquake mechanism and appears characteristic of areas susceptable to induced seismic activity in the southeastern United States. Further development of the ω^{-3} discriminant in the southeastern United States and the application to other areas could provide a method for prediction of seismic acitivity at a given reservoir site.

The success of the application of such a spectral discriminant depends primarily on obtaining data for which the spectral slope can be reliably computed. An analysis of the limitations induced by the recording system and frequency dependent attenuation was given in the semi-annual technical report. The major difficulty encountered in this research was the sparsity of data satisfying the conditions required for a successful computation of the high-frequency slope. Seismograms from instruments with well-defined response at frequencies above the corner frequency are needed to determine the high-frequency slope. The limited high-frequency response of most systems does not allow computation of the spectral slope past the corner frequency for small events ($M_L<1.0$) and larger events may be saturated. At greater distances the high-frequency spectra are lost to anelastic attenuation. The single largest problem in relating spectral characteristics to earthquake source parameters is accounting for anelastic attenuation between source and station. Seismic data must be chosen so that sufficient information at frequencies above the corner frequency is retained.

A high dynamic range digital system recording at short epicentral distance is optimum for spectral information as related to the source. Telemetry systems and portable high-frequency magnetic tape recorders were found to record reliable earthquake spectral data under certain conditions.

The lack of appropriate instrumentation has eliminated the possibility of obtaining recordings of induced earthquakes from most reservoir areas. Cooperation was asked in obtaining data from selected reservoir areas outside the southeastern United States for which suitable recordings might be available. Very little appropriate data was located and none was found which could reasonably (either logistically or within our time constraints) be released which could be reliably analyzed for high-frequency spectral characteristics. Therefore, this research has of necessity been limited to examination of additional southeastern United States data and published spectral data from two reservoir areas in the western United States.

II. Induced Seismicity Discrimination Efforts

The problem of determining the relationship between induced and natural earthquakes at reservoirs and other areas where a non-natural mechanism for inducing earthquakes exist has been the subject of several studies. Accumulated evidence directly supports a casual relation between loading a reservoir and the ensuing earthquakes and between injection of fluid and earthquakes.

Investigations of several cases of reservoir induced seismic activity show several common characteristics (Gupta and Rastogia, 1976). Deep or very large reservoirs have instigated a higher occurrence of reservoir induced seismicity. The largest reservoir related earthquakes are all about magnitude 6. The number of reservoir associated events increase as the lake level rises, with the largest earthquakes occurring after the lake has reached its highest level. Variations in lake level are often accompanied by variations in local seismicity. The b values for reservoir associated earthquake sequences are usually higher than those of normal sequences. In all cases of reservoir induced seismicity, the rocks within which the earthquakes occurred were presumed to be highly stressed before impoundment even if no contemparenous activity had been noted (Kisslinger, 1976).

The number of large reservoirs is rapidly increasing with over 50 deep and/or very large reservoirs expected by 1985. A curve showing the occurrence of reservoir induced seismicity for deep or very large reservoirs over the past 20 years suggests that about 10 new cases of reservoir induced seismicity are likely to occur (Packer et al., 1978). Statistical analysis has been used in many attempts at correlating induced seismicity levels with reservoir parameters. By sampling statistical variances for parameters of maximum water depth, reservoir volume, stress state of rocks, recent fault activity, and surface geologic units in areas of reservoir induced seismicity, Packer et al. (1978) were able to define the probability of seismicity being caused by large reservoirs. Their analysis developed data with which the conditional probability of reservoir induced seismicity, given any of the five specific states, could be calculated. Reservoir depth was found to be the most distinguishing attribute among the five parameters. Reservoir volume was also found to be an important factor perhaps because it relates to depth. No other significant relations were observed. A statistical correlation was attempted between the frequency of seismic activity at reservoir sites and the seismic activity of the surrounding region (Daly et al., 1977). In no case did the seismic activity near reservoir areas equal or exceed the predicted seismicity. However, the survey data and predicted values used were approximate and the amount of error was unknown. The probability that a large reservoir would induce seismic activity was calculated to be 0.007.

An experiment was conducted by Peppin and Bufe (1979) to determine if events occurring near the Geysers Geothermal area, California could be seismically discriminated from events near the San Andreas Fault system. Focal mechanisms, spectral corner frequency, seismic moment, and Richter magnitude were compared for the two groups of earthquakes. These parameters did not allow a discrimination between the two sets of data. Their result was qualified in that the geyser events, may have been naturally occurring. Also, in our opinion the conclusions reached could be inappropriate to reservoir induced earthquakes because of different stresses or triggering mechanisms and the potentially large thermal stresses involved.

III. Theoretical Basis for a Reservoir Induced Earthquake discriminant

Seismic energy radiated from the earthquake focus carries frequency and amplitude information which can be related to the source character. Farfield body-wave spectral amplitudes have been used to test various source models and to evaluate their characteristic parameters (Haskell, 1964, 1966; Aki, 1967; Brune, 1970, 1971; Savage, 1972; Dahlen, 1974; Madariaga, 1976). Most models of the seismic source are based on the concept of relaxation of stress. These are usually developed in terms of either a tangential shear dislocation (Brune, 1970; Hanks and Wyss, 1972) or a longitudinal shear dislocation model (Haskell, 1964; Niazi, 1974).

Stress relaxation of the source predicts certain properties of the displacement spectra. The far field spectra are constant at low frequencies and inversely proportional to some power of frequency at high frequencies. The transitions may be sharp or gradual indicating one or more corner frequencies. The static parameters, seismic moment and fault dimensions are inferred from the extrapolated zero frequency intercept and the frequency of the first corner (Aki, 1967; Brune, 1970; Savage, 1972). The displacement spectra are flat at low frequencies because the source is effectively a point source for long wave lengths (Molnar et al., 1973). The maximum spectral amplitudes occur at zero frequency and must decrease in amplitude with increasing frequency so as to have finite energy. Most source models define the high-frequency spectral slope of the Fourier transform of the source time function with an asymptotic decay proportional to ω^{-n} where n>1.5 (Molnar et al., 1973). The corner frequency, defined as being intermediate to the low and high frequency trends, is inversely proportional to the pulse width of the displacement signal. The relationship between the spectral corner frequency, and the model parameters depends on the geometrical design of the assumed source model.

The ratio of P- to S-wave corner frequency is usually found to be >1.0 (Molnar et al., 1973; Hanks and Wyss, 1972; Molnar and Wyss, 1972; Marion, 1977; Rastogi and Singh, 1977; Rautian et al., 1978). Some observations have found S-waves to have higher corner frequencies than P-waves (Bakun et al., 1976; Peppin and Simila, 1976). A directed high rupture velocity has been postulated to cause a shift in the S-wave spectra such that $f_p > f_s$ (Savage, 1972). Theoretical work by Weertman (1971, 1975), Savage (1974), Dahlen (1974), and Burridge (1975) has shown that for a transonic rupture (rupture velocity greater than S-wave velocity), the P-wave corner frequency is greater than the S-wave corner frequency over most azimuths.

In a study of high frequency seismic radidation, Madariaga (1977) found P waves to be only slightly affected by rupture velocity. The SH and SV waves radiated slower, with consequently lower frequencies, at higher fault rupture velocities due to a focusing effect. The radiation of P-waves was found to be more efficient at higher frequencies.

Therefore, P-wave corner frequencies should usually be higher than those of S-waves for higher rupture velocities. The radiation of high frequencies is controlled to a large extent by the slip velocity on the fault. Velocity changes on the rupture front cause release of high-frequency radiation when the front accelerates or decelerates. The upper bound to the high-frequency radiation is determined by abrupt changes in rupture velocity as caused, for example, by stopping phases. When the rupture velocity is constant, changes occur only due to a variation in stress intensity at the rupture front and there is subdued high-frequency radiation. Strong high-frequency radiation occurs when the rupture velocity changes abruptly. In this case, higher frequency SH and SV waves are radiated and $f_p/f_s \leq 1.0$.

In dislocation or crack (stress drop) models, the high-frequency radiation is determined by starting and stopping phases. For these models, crack extension causes stress concentration to become larger causing an increase in rupture velocity. After the initiation of dislocation, the rupture velocity tends to accelerate up to the shear wave velocity for antiplane cracks (Kostrov, 1966) or the Rayleigh velocity for inplane shear cracks (Fossum and Freund, 1975). Transonic rupture velocities have been proposed by Burridge (1973), Andrews (1976b), and Das (1976). The mechanism of failure has been modeled as either a fracture or a strick-slip function (Andrews, 1976a). Weertman (1969) calculated the velocity at which earthquake dislocations should propagate if the friction law of a fault is dominated by a dependence on the slippage velocity. The dislocation traveled at the Rayleigh velocity (edge dislocations) or the shear-wave velocity (screw dislocations) or at a supersonic velocity.

Numerical calculations were used (Andrews, 1976b) to show how the transition between sub-Rayleigh and super-Rayleigh rupture propagation depends on the three parameters; stress limit, fracture surface energy, and crack length. The speed of rupture can abruptly change when the rupture reaches differing stress or frictional regimes, which are controlled by variations in the local geology (Achenbach and Harris, 1978). As expected, the high-frequency motions are radiated at abrupt changes of the rupture front speed. A discontinuous rupture front speed on a singular front produces more severe high-frequency motions than would be produced by a gradual change of the speed in conjunction with a transition zone for rupture (Husseini and Randall, 1976). In the frequency spectrum, the frequency range higher than the corner frequency is dominated by the high frequencies generated by a discontinuous rupture front speed. These spectra decay slower than spectra dominated by a gradual change of rupture velocity.

Madariaga (1977) determined that starting and stopping phases provide strong upper bounds to the radiation of high frequencies and a ω^{-2} high-frequency dependence is the slowest possible spectral decay at high frequencies. When the rupture velocity changes are smooth ω^{-2} is expected to be the upper bound of the high frequency spectra with ω^{-3} being more common.

The shape of the displacement spectra was found by Savage (1974) to be determined by the empirical relation $\xi = v \sin\theta / c$ where v is the rupture velocity, θ is the angle between the direction of wave propagation and the normal to the fault, and c is the wave velocity.

Farfield amplitude spectra for subsonic and transonic models of P and SH waves for 3 source models were computed by Niazi (1974). The subsonic models (2.5 km/s) show more high frequencies generated and a spectral slope which has less decay than the transonic (5.0 km/s) earthquake spectral models.

For a 3-d fault whose rupture nucleates in a small localized zone, Dahlen (1974) has shown starting phases will have an ω^{-3} decay instead of the ω^{-2} of line sources. In most simple models of constant rupture velocity where the rupture stops abruptly at a barrier, the strongest high-frequency radiation is associated with stopping phases (Savage, 1966; Madariaga, 1976) which can cause a ω^{-2} decay.

Overall, the theoretical models indicate that subsonic rupture velocities on a fault with irregular tractional resistance give spectra with multiple corner frequencies and ω^{-2} or less decay at high frequencies. Smooth faults with tractional resistance less than 5.32 times the driving shear stress (Fossum and Freund, 1975) allow transonic rupture velocities and generate spectra with ω-cube decay.

IV. Observed Spectral Data

The Clark Hill and Jocassee Reservoir areas, located in the Piedmont Province of the southeast United States, are considered to be examples of induced seismic activity. An ω^{-3} high-frequency amplitude decay was observed for earthquakes located near these reservoirs (Marion, 1977). A tensional deviatonic stress environment, conducive to normal or strike-slip faulting is determined for these reservoir areas (Guinn, 1977; Talwani *et al.*, 1978). Earthquakes from near Maryville, Tennessee, located in the Folded Appalachian Province where reservoirs have not induced seismic activity, typically show an ω^{-2} high frequency decay. This area is generally believed to be in a compressive stress environment.

The literature was surveyed in the hope of identifying other areas characterized by an ω^{-2} or ω^{-3} spectral decay. Table 1 lists observed published spectral data as to author, area, and spectral characteristics. Most published spectra data are for California earthquakes. These generally show an ω^{-2} decay for events of southern California, where compressive stresses dominate. Reservoirs in southern California have not triggered earthquakes. Oroville reservoir, in the eastern California Sierra Nevada foothills, is the only California reservoir considered to have induced seismic activity. A magnitude (ML) 5.7 earthquake occurred in an area of relatively low historical seismicity 10 km from Lake Oroville on August 1, 1975. Displacement spectra of aftershocks of this event were calculated by Fletcher (1979). The high-frequency fall-off rate of these events was greater than ω^{-2} and frequently as high as ω^{-4} or ω^{-5}. The Oroville reservoir area is considered an area of dominantly normal or strike-slip faulting (Bufe _et al_., 1976; Ryall _et al_., 1976; Bell and Nur, 1978).

Studies of focal mechanisms and source parameters at Lake Mead, Nevada-Arizona were made by Rogers and Gallanthine (1973) and Rogers and Lee (1976). A right-lateral strike slip motion on vertical fault planes was generally observed, with a possibility of normal faulting. Source parameters obtained for a group of seven Lake Mead events (S-waves) showed a high-frequency decay of $\omega^{-1.3}$ to $\omega^{-1.8}$ (Rogers and Gallanthine, 1973). They concluded that tectonic forces and not the reservoir were chiefly responsible for these events.

Spectra of some Himalayan earthquakes were calculated by Singh _et_

Table la. Observed Published spectral data

Author	Area	Spectral Characteristics
Aki and Chouet (1975)	Stone Canyon, CA near San Andreas	ω^{-2} high frequency decay of coda waves
Bakun and Bufe (1975)	Bear Valley, CA	S-wave spectra $\omega^{-1.5}$ to ω^{-2} at > 2 Hz, fc for 1.1 $\leq M \leq 2.2$ at >10-12 Hz
Bakun *et al.* (1976)	Central CA	frequency $\geq$10 Hz SH spectra decrease more rapidly than PZ spectra. Q_p = 175-250 Qs = 100-150, ω^{-2} to ω^{-4} high frequency decay
Bakun *et al.* (1978)	San Andreas Fault, CA	No slope given
Frasier and North (1978)	Rat Island, South-west of Alaska	ω^{-3} high frequency spectral decay
Hanks and Wyss (1972)	CA, Iran, Turkey	Interpreted spectral data to decay as ω^{-2} at high frequencies
Johnson and McEvilly (1974)	Central CA	ω^{-2} to ω^{-3} high frequency decay
Peppin (1976)	Nevada Test Site	explosion data decay $\geq \omega^{-3}$ higher corner frequency than earthquakes
Peppin and Simila (1976)	Trans Sierra-Nevada CA-NEV	P and SV spectra show ω^{-2} or ω^{-3} high frequency decay $Q_s \geq 480$
Ryall *et al.* (1976)	Oroville Reservoir, CA	P and S-wave spectra ω^{-2} to ω^{-3} decay - Mag 3.0-4.3, corner frequencies 10-20 Hz
Tanis (1973)	Worldwide earthquakes	Coda displacement spectra slope not found
Thatcher and Hanks (1973)	Southern CA	ω^{-2} high frequency decay close in, $\omega^{-1.5}$ at distant stations
Wyss *et al.* (1971)	Aleutian Islands	Spectral data 0.5 $\leq T \leq$33 sec. ω^{-2} for earthquakes and explosions.

Table 1b.

Author	Area	Spectral Characteristics
Fletcher (1979)	Oroville, CA	High-frequency spectral decay $\geq \omega^{-2}$ and frequently as high as ω^{-4} for 10 Hz $c \leq fc \leq 70$ Hz.
Molnar *et al.* (1973)	San Fernando, CA	fp>fs, spectral slope not determined
Singh *et al.* (1979)	Himalayas	earthquakes in thrust environment typically $< \omega^{-2}$ decay
Thatcher (1972)	Northern Baja, CA	earthquakes $> M_L = 3.7$ generally $\leq \omega^{-2}$
Rogers and Gallanthine (1973)	Lake Mead	found $\omega^{-1.3}$ to $\omega^{-1.8}$ for S-wave spectra
Wyss and Hanks (1972)	San Fernando, CA	S-wave spectra decay as $\omega^{-1.5}$ P-wave spectra decay as $\omega^{-1.8}$
Wyss and Shamey (1975)	Tonga Islands Kamchatka Islands	Long period spectra No slope found

al. (1979). These events, in a thrust fault environment, typically showed a spectral decay of $< \omega^{-2}$. Tarbela Reservoir, also in the Himalayas, on the Indus River in northern Pakistan is in an area of considerable natural seismicity. The local seismicity, under compressional stress, has been shown to decrease following reservoir loading (Jacob et al., 1979).

Published spectral data have shown an ω^{-3} decay in reservoir areas which have induced seismic activity. These reservoirs are typcially characterized by normal or strike-slip faulting. Of the major cases (magnitude $\geq$ 6.0) of reservoir induced seismicity, strike-slip or normal fault plane solutions have been obtained (Simpson, 1976). Reservoirs located in areas of compressive stress (Folded Appalachian Province, Himalayas) have had no effect on or lessened local seismic activity. In regions where thrust faulting dominates, stress due to a vertical load should have a minimum effect on local seismicity (Simpson, 1976; Jacob et al., 1979).

V. Propagation Effects and Recording Limitations

Limitations on reliable high-frequency spectral computations are imposed by frequency-dependent absorptive attenuation which acts as a low pass filter for seismic waves in the earth.

Seismic recording systems with a narrow-band response and low dynamic range selectively reduce the information amenable to spectral analysis. Small earthquakes can have corner frequencies which exceed the upper frequency limit of the recording system and therefore the spectral slope cannot be computed. Also, saturation at high gain can distort the frequency content of larger magnitude earthquakes.

In a previous report, Long and Johnston (1979) examined earthquake magnitudes in comparison to corner frequency, theoretical Q values, and seismic system frequency response and gain to calculate appropriate data for high frequency analysis. Appropriate parameters for reliable recordings of discrete earthquakes were calculated for WWSSN and for a typical telemetry system. Figure 1 shows the results of analysis for a typical telemetry system. Recordings of earthquakes of magnitude 2.0 ± 0.3 at distances of 20 to 50 km in regions where Q is greater than 400 are found to be suitable. An increase in Q extends the reliable high frequency data range. Figure 2 shows shows appropriate WWSSN data. The region (of reliable data) is so small that we conclude that most WWSSN data are not appropriate for high frequency spectral analysis to measure the slope past the corner frequency.

A high frequency (15-400 Hz) magnetic tape system has been used to record microearthquakes for spectral analysis in the southeastern United States (Marion, 1977). These microearthquakes, with frequencies in the range of 30-200 Hz, were recorded at epicentral distances of less than 4 km. High dynamic range digital systems have also successfully recorded earthquakes at similar near distances (Tucker and Brune, 1972; Fletcher, 1979). These systems are ideal for calculation of the high frequency spectral slope as the high-frequency information out to about four times the corner frequency are readily available.

VI Data Regions

The success of this study was dependent on obtaining earthquake data from as many reservoir areas as possible. Data deficiencies have limited these analyses because most reservoir areas were not monitored with seismic instrumentation prior to filling and the few reservoirs

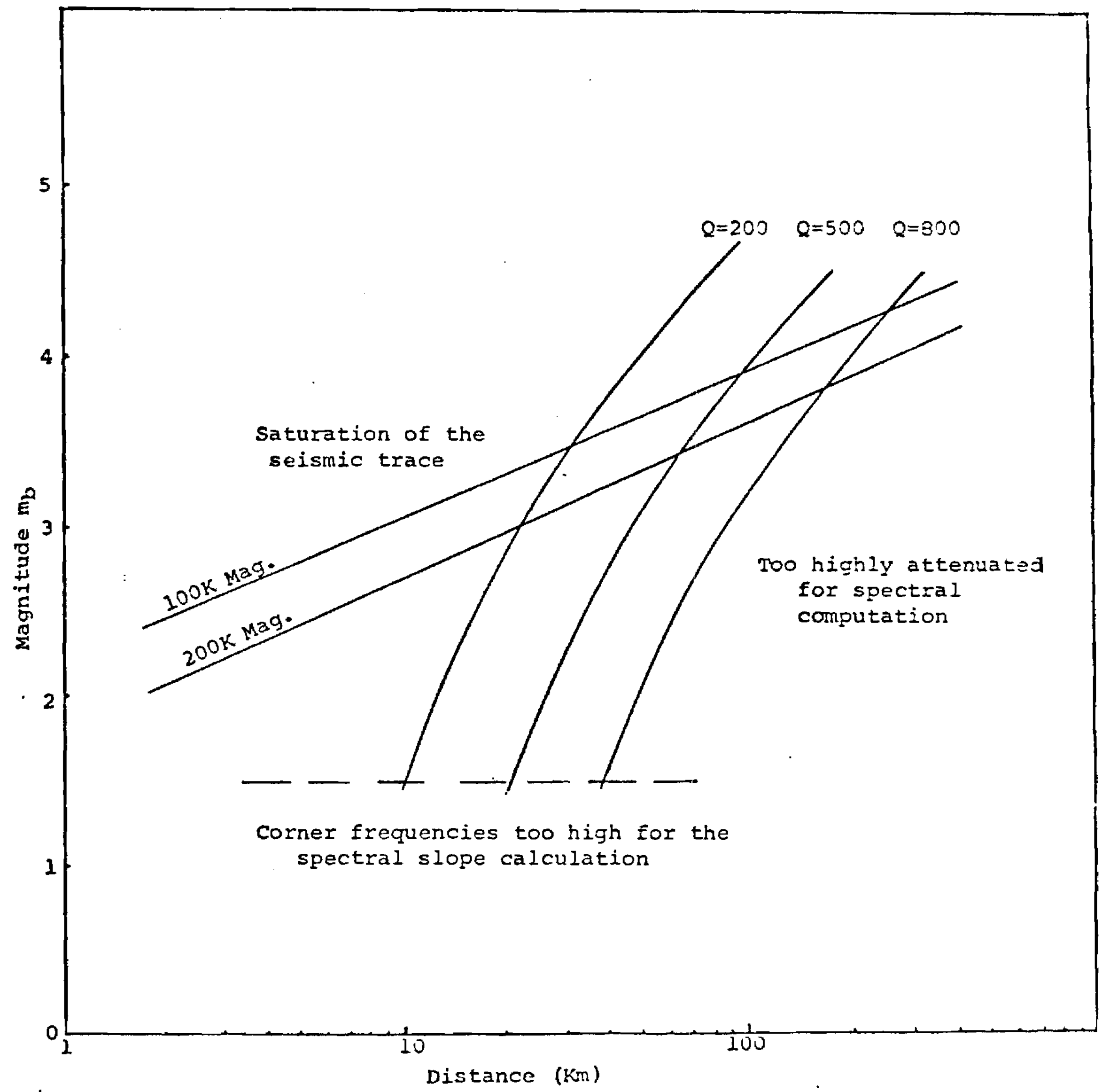

Figure 1. Range of magnitudes versus distance that can be used for determination of the spectral slope past the corner frequency for a typical telemetry system using Nuttli's m_b.

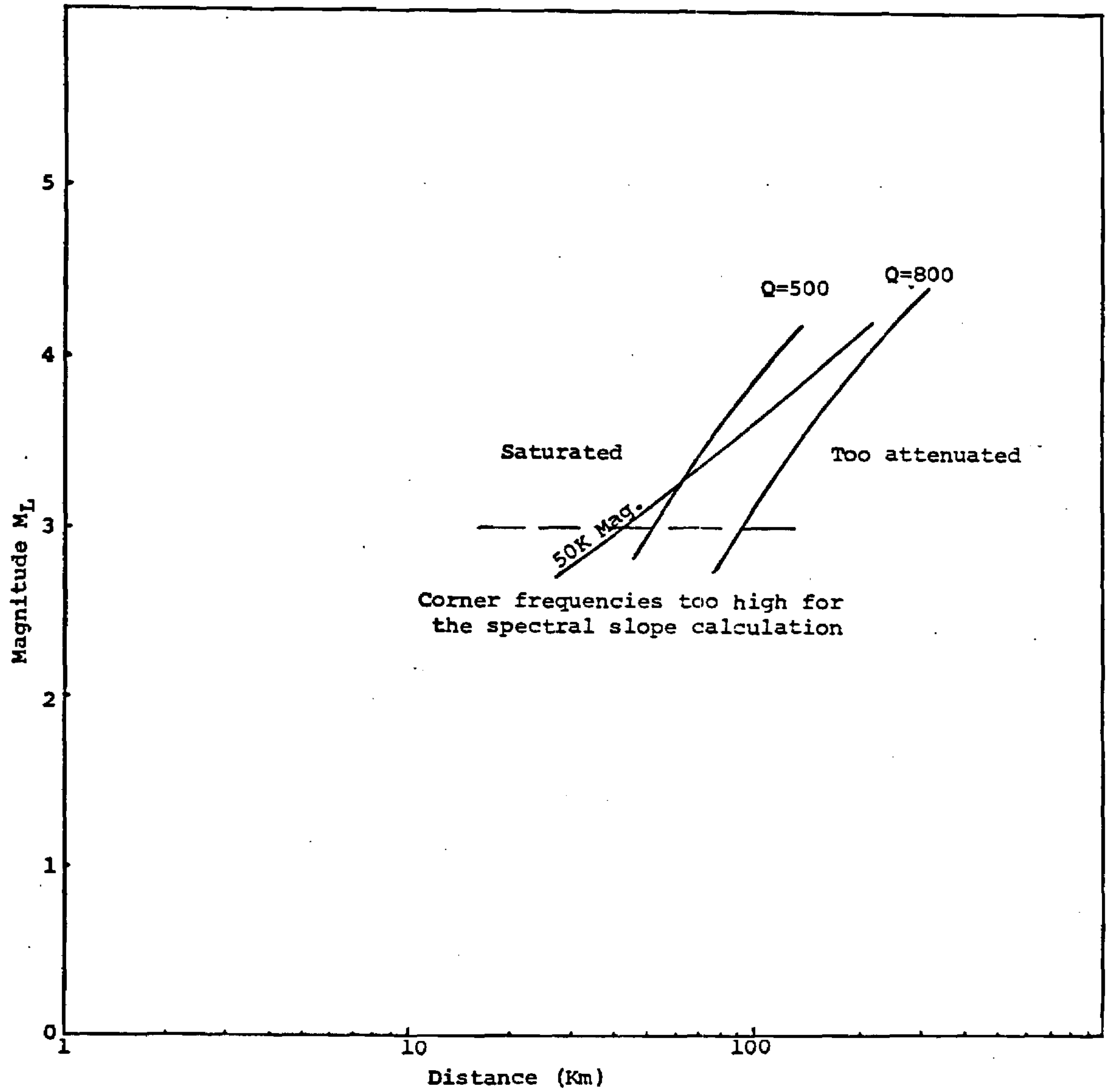

Figure 2. Range of magnitudes versus distance that can be used for determination of the spectral slope past the f_c for WWSSN SP using Richter's M_L.

that have been monitored do not generally provide adequate data for spectral analysis. Also, naturally occurring earthquakes near reservoir areas are often not reported or not found. In the semi-annual technical report, possible reservoir data areas were tabulated. Of these, most were inadequately instrumented or data were not available. Reservoir areas for which data are currently available or may be available are shown in table 2.

Table 2. Reservoir areas for which spectral data may be available.

Reservoir	Reference	Data type
Clark Hill (GA-SC)	Marion, (1977)	magnetic tape
Jocassee (SC)	Marion, (1977)	magnetic tape
Monticello (SC)	(this report)	magnetic tape
	Fletcher, Talwani, Duc (in progress)	digital
Lake Sinclair (GA)	(this report)	magnetic tape
Oroville (CA)	Fletcher (1979)	digital
	Ryall _et al_., (1976)	
Southern Cal. Res. (CA)	Daly _et al_., (1977)	unknown
Lake Mead (NV)	Rogers & Gallanthine (1973)	digital
Manic 3, (Canada)	Leblanc & Anglin (pers. comm. 1979)	magnetic tape

VII. Tests of Spectral Computation Techniques

New data were obtained during the course of this research on magnetic tape recorders described by Marion (1977). Data recorded on magnetic tape were played back on a strip chart recorder at speeds up to 125 mm/sec. The strip chart record was photographed and then magnified to insure an accurate and convenient trace for digitization. The traces were then digitized using an analog to digital graphical on-line converter and prepared for spectral computation.

We have examined the influence of several factors which can degrade the Fourier frequency spectra. Aliasing, the conversion of energy in frequencies above the folding frequency to energy in low frequencies in the spectra, occurs when the digitizing interval is not small enough to define all the character of the trace. Wave forms were digitized with at least 8 and usually 10 or more points per half cycle. The Nyquist (or folding) frequency, above which aliasing occurs, is defined as being one-half the digitizing rate (Kanasewich, 1975). The digitizing rate used on data in this report gives a folding frequency which is above the response of the recording system so that aliasing did not introduce errors in the spectra. Clipping of the time trace can effectively cause a distortion in the high-frequency spectra and increase noise levels. A spectra of a saturated trace is shown in Fig. (3). Geophone placement or seismic system problems can generate distorted spectra. Figure (4) show spectra of a resonating time trace caused by a resonating geophone. Note a dominant spike at a resonance frequency (20Hz). Data with the above problems were eliminated from analysis. Noise can adversely affect the calculated spectra. Noise at the station site effectively reduces the usable dynamic range of the earthquake time trace. The effects of digitizing noise on a displacement spectra have been calculated by two methods. A cosine function was Fourier analyzed using the same techniques as were used in digitizing the seismograms to obtain a theoretical noise spectra. Fig. 5 shows the spectra obtained. Actual noise spectra were calculated for a portion of the time trace immediately preceding arrival of the P phase for some earthquakes. Fig. (6) shows a typical seismic noise spectra. The digitizing noise level increased with an increase in the trace amplitude. Since the digitizing noise was virtually a white noise spectra the instrument corrected noise spectra correlates with the inverse of the frequency response curve of the seismic system.

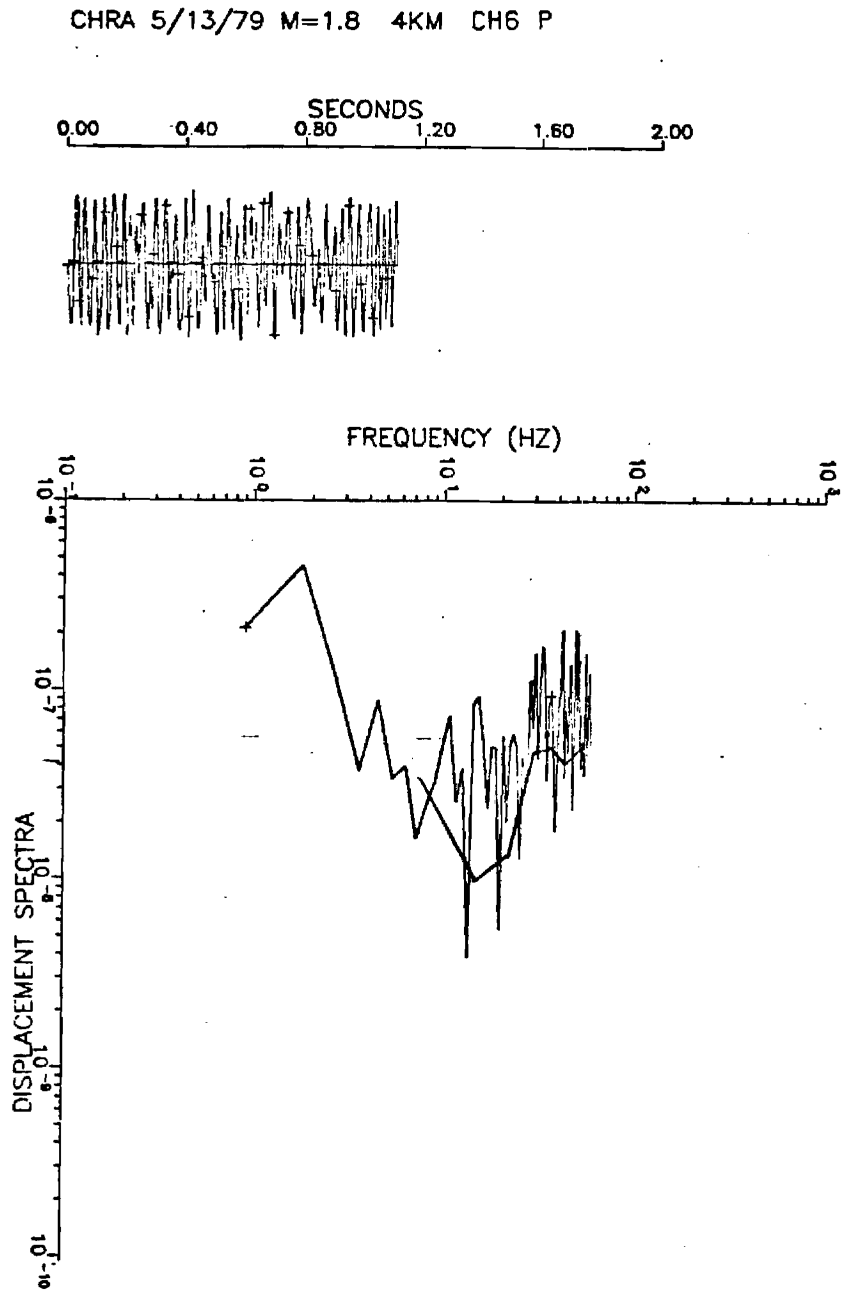

Figure 3. Spectrum of a seismic trace for a event large enough to saturate the recording system. In this example the source spectra is masked by noise.

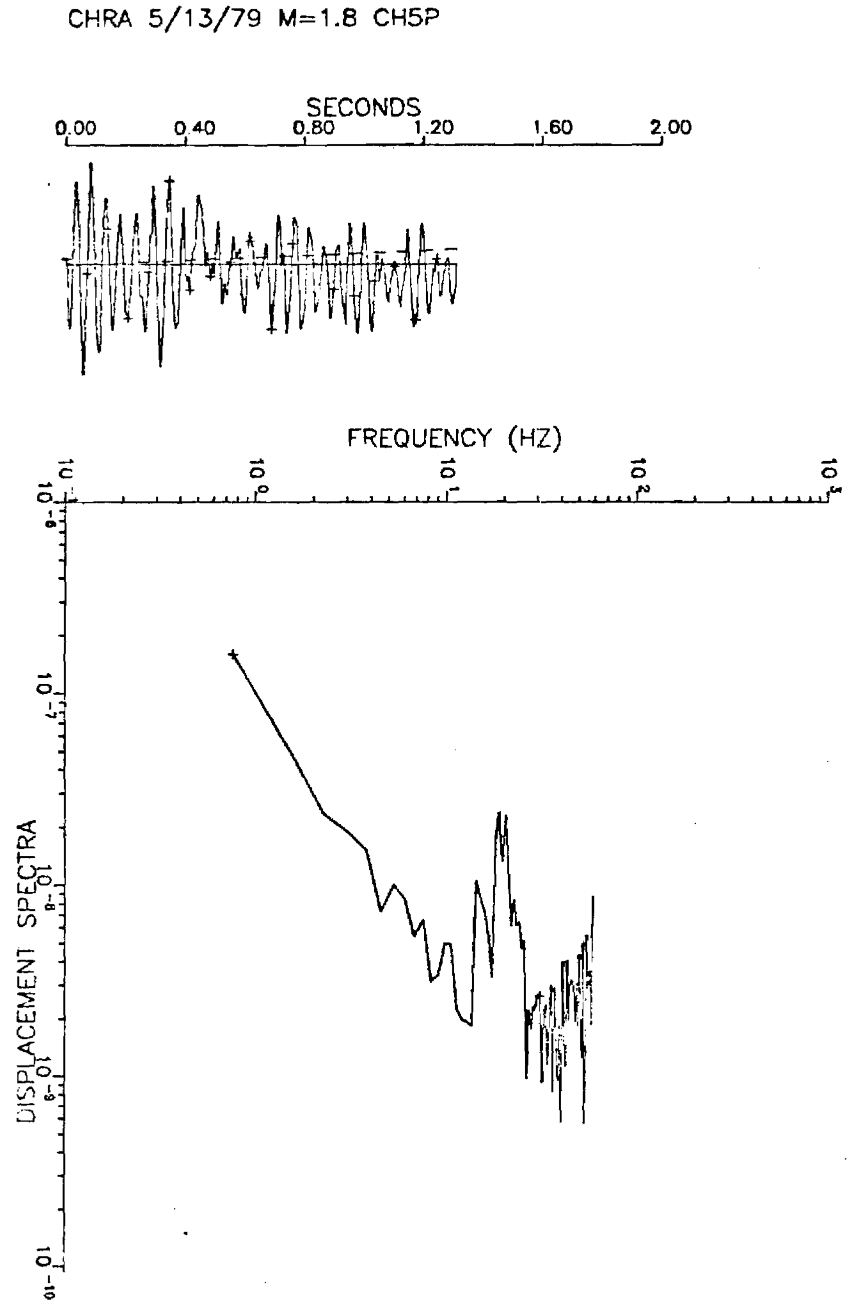

Figure 4. Spectrum of an earthquake recorded on a system with a spurious resonance at about 20 Hertz.

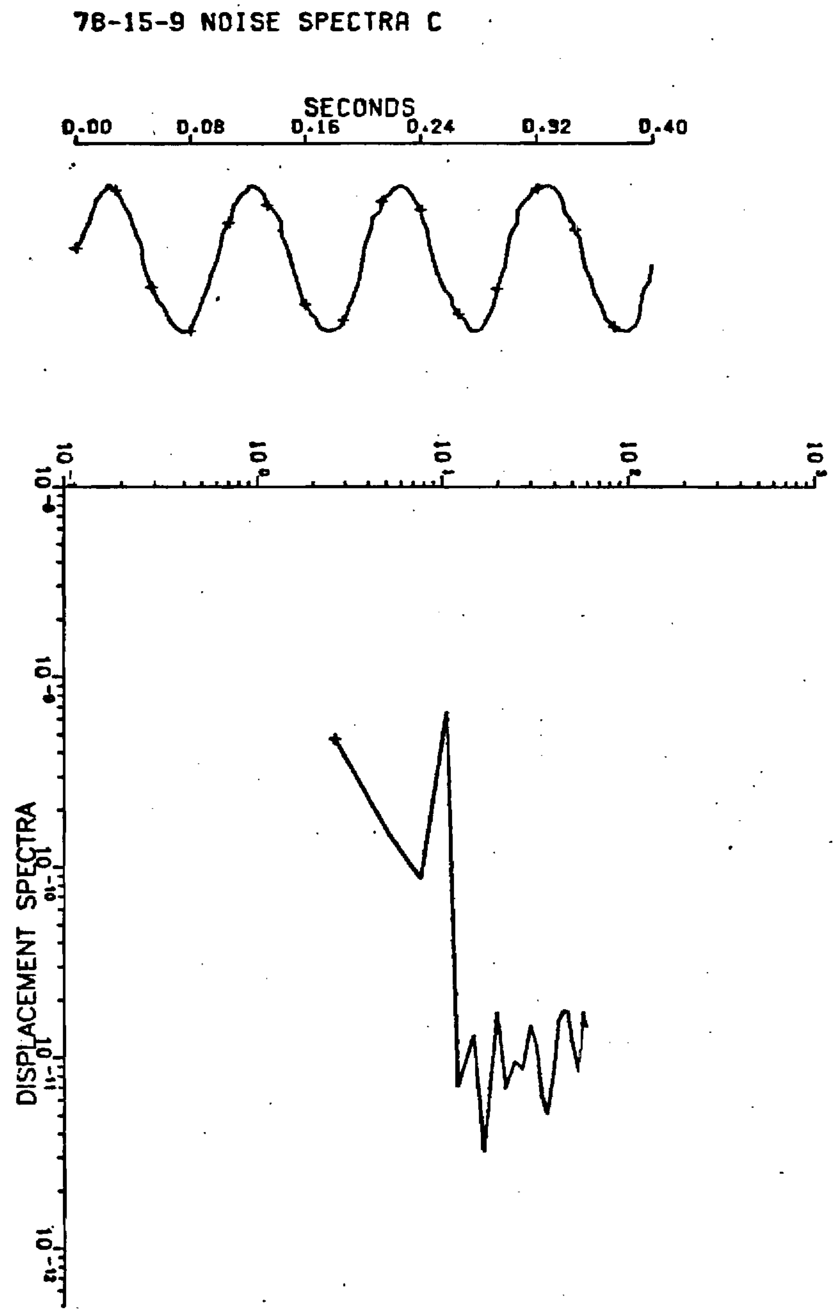

Figure 5. Spectrum of a Cosine function used to study magnitudes of digitizing noise when a trace is present.

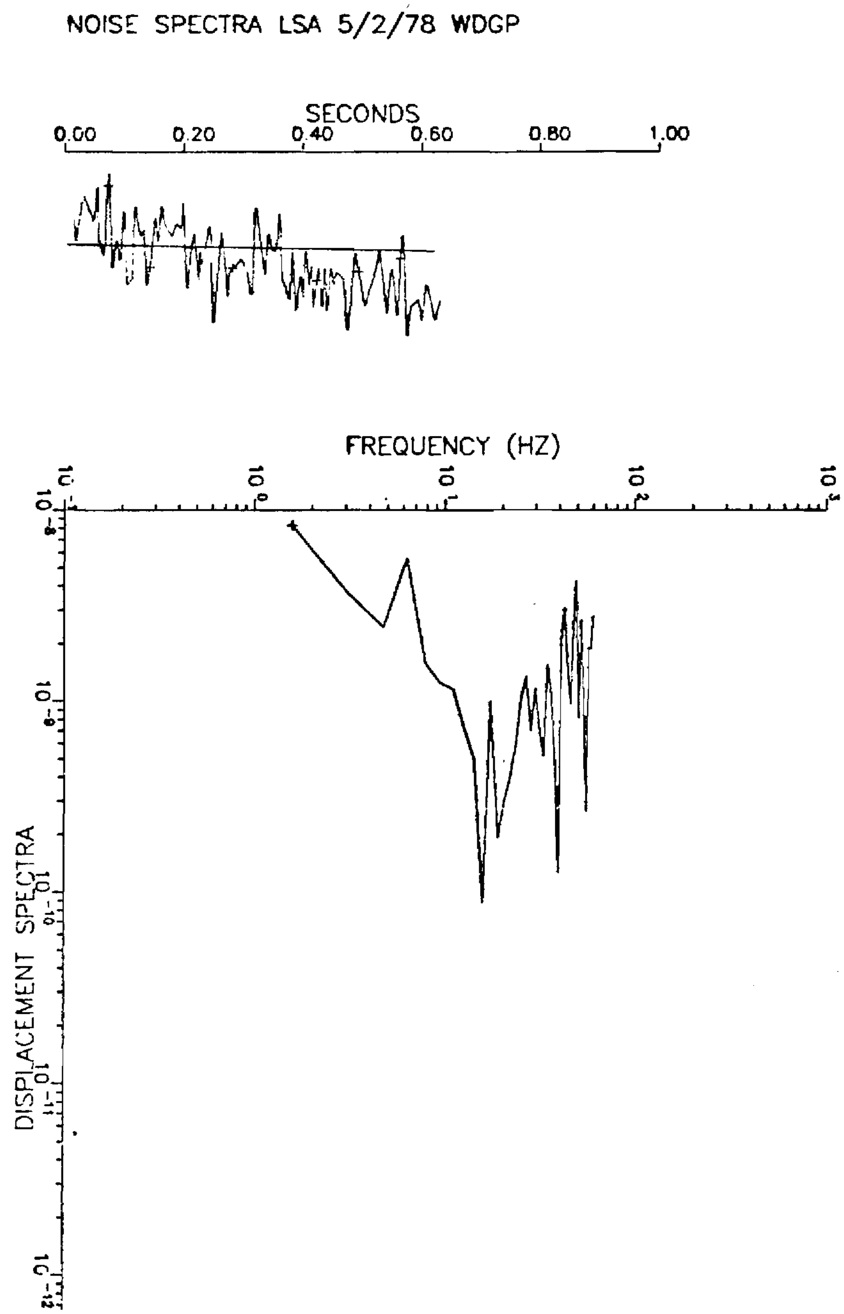

Figure 6. Seismic noise preceding the P-wave arrival of a magnitude 2.3 Lake Sinclair area event.

Spectra were calculated using a Fast Fourier transform algorithm. Results are presented in a plot of log-displacement amplitude versus log-frequency. A plot of the earthquake time trace for which the spectra were computed is also shown above each plot. P- and S-waves were chosen for analysis such that as much reliable information as possible was retained. The seismogram trace is corrected for base line shift by fitting a least squares best-fit straight line to the digitized data. A base line through the data is subtracted from the amplitude of each point. The data are corrected for the displacement response of the total system in the frequency domain. No tapers were applied to smooth the edges of the data. Fig. (7) shows a typical displacement spectra. To determine the effects caused by discontinuities at the ends of the time window the trace of Fig. (7a) was also padded with zeros at both ends and at one end before computing the Fourier transform. The resultant spectra (Fig. 7b and 7c) are virtually identical to Fig. 7. Truncation has its most severe effect when the traces contain less than about two cycles. A Bartlett spectral estimate was used to reduce the variance of some of the spectra. The time trace is divided into a discrete even number of overlapping segments. Each segment was windowed and transformed to the frequency domain where the segments were averaged and the effects of the window were removed (Kanasewich, 1975, pg. 99). The resulting spectrum has a variance reduced proportional to the inverse of the number of segments and can be used to more accurately predict corner frequencies and high frequency slopes if the original time trace is sufficiently long. Fig. (7d) shows the Bartlett spectral estimate of the spectra of Fig. 7a.

Maximum entropy spectral analysis (Burg, 1972) has been considered

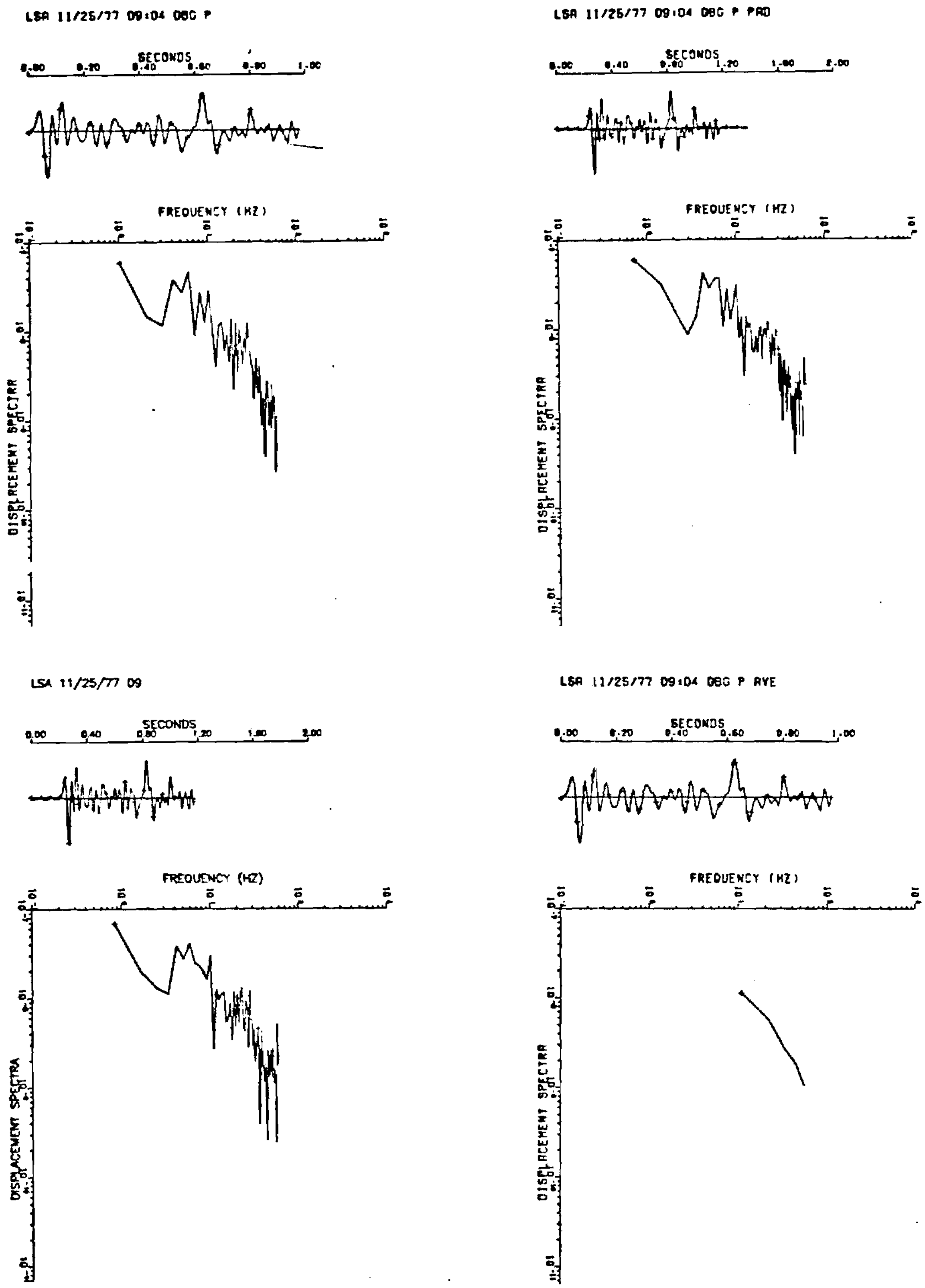

Figure 7. Typical displacement spectrum used to show the influence of padding the trace with zeros. (a) No zero padding as used on other spectra in this report. (b) Zero padding in front and back of trace. (c) Zero padding in front only. (d) Bartlett spectral estimate of the same trace showing reduction in variance.

as a possible means of spectral power estimation for short (i.e. 2 to 4 cycles in the time trace) data segments where Fourier analysis may not be appropriate. Maximum entropy spectral analysis is used typically to obtain power spectra when the signal is short or periodic and not completely sampled. The Fourier transform has poor resolution in analyzing sinusoidal components which extend beyond the data window. As the earthquake time traces were usually of sufficient length and varying frequency, we found no advantages and some disadvantages in using Maximum Entropy Spectral Analysis. For our data the Fast Fourier Transform or cosine transform is preferable because it was simpler to apply.

VIII New Data Analyses

Clark Hill Reservoir Area (CHRA). The Clark Hill Reservoir is located along the Georgia-South Carolina border on the Savannah River approximately 50 km north of Augusta, Georgia. The northern portion of the reservoir was the site of the August 2, 1974 magnitude 4.3 earthquake. The CHRA is located in the crystalline Piedmont Province. A study of microearthquake displacement spectra of earthquakes in the epicentral area of the August 2, 1974 earthquake was iniated by Marion (1977). An ω^{-3} high-frequency spectral slope was observed for most earthquakes studied. Magnetic tape data obtained by Guinn (1977) was used to compute focal mechanism solutions for microearthquakes which occurred on March 26, 1977 and April 14, 1977. The solutions obtained were similar to that of the August 2, 1974 main event. The focal mechanism solutions indicate movement on nearly vertical faults with no uniform preference for normal or reverse type motion. Station and epicentral locations of Guinn's data are shown in Figure (8). These

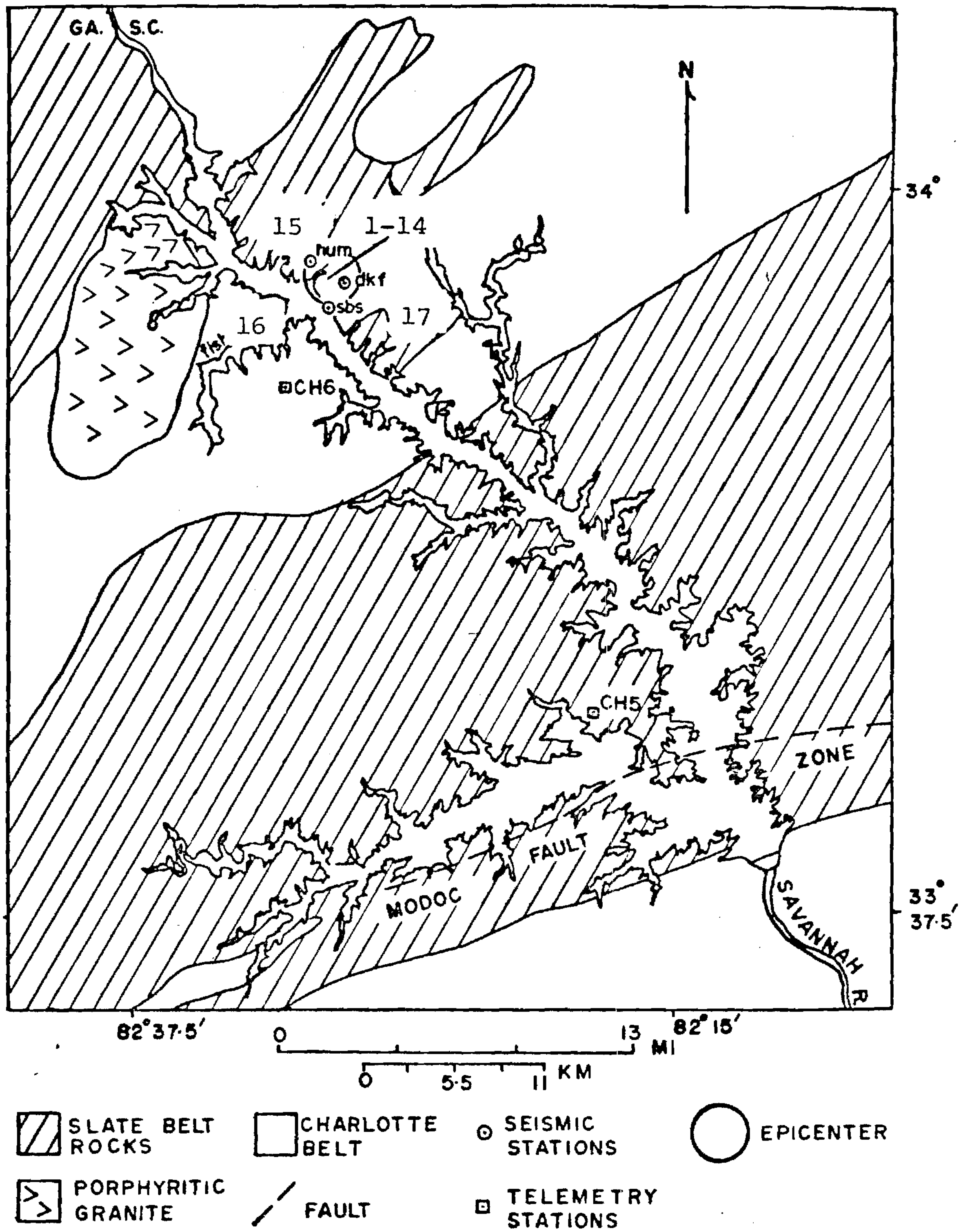

Figure 8. Clark Hill Reservoir epicentral area with seismic stations and earthquake locations.

data were reexamined for their spectral characteristics. Twenty-two P and S-wave spectra were obtained. Fig. (9) shows a typical CHRA microearthquake P-wave spectra. Table (3) lists magnitude, corner frequency, and high-frequency decay rates of these events. The slopes decay essentially as ω^{-3} with only a few decaying as ω^{-2}. The ratio fp/fs for these events was generally greater than one.

Georgia Tech telemetry system data has been used to calculate the slopes of a few larger CHRA events. These recordings fit the criteria established for reliable spectral analysis above the corner frequencies. All telemetry data showed a ω^{-3} decay.

Monticello Reservoir Area (MRA). Monticello Reservoir is located in Fairfield County, South Carolina about 45 km northwest of Columbia. It is the source of cooling water for the Virgil County Summer Nuclear Station. The reservoir was filled in December, 1977 and seismic activity started immediately. Seismic energy release, above the pre-impoundment level, has continued to the present. The Monticello Reservoir Area (MRA) is located within the Charlotte Belt metamorphic zone of the Piedmont Province. The area is underlain by almandine-amphibolite facies metamorphic rocks which have been intruded by plutons of granite to granodioite composition. Faulting has been observed to have occurred in the area. However, major movements probably took place 150-300 m.y. ago (Talwani et al., 1978).

The steady rate of minor activity and easy availability of data at Monticello Reservoir prompted field studies in the area for the purpose of recording data which could be analyzed for high-frequency spectral decay. Arrays of smoked paper and high frequency magnetic tape recorders were used to record several small (mag $\leq$ 0.0) events. Fig.

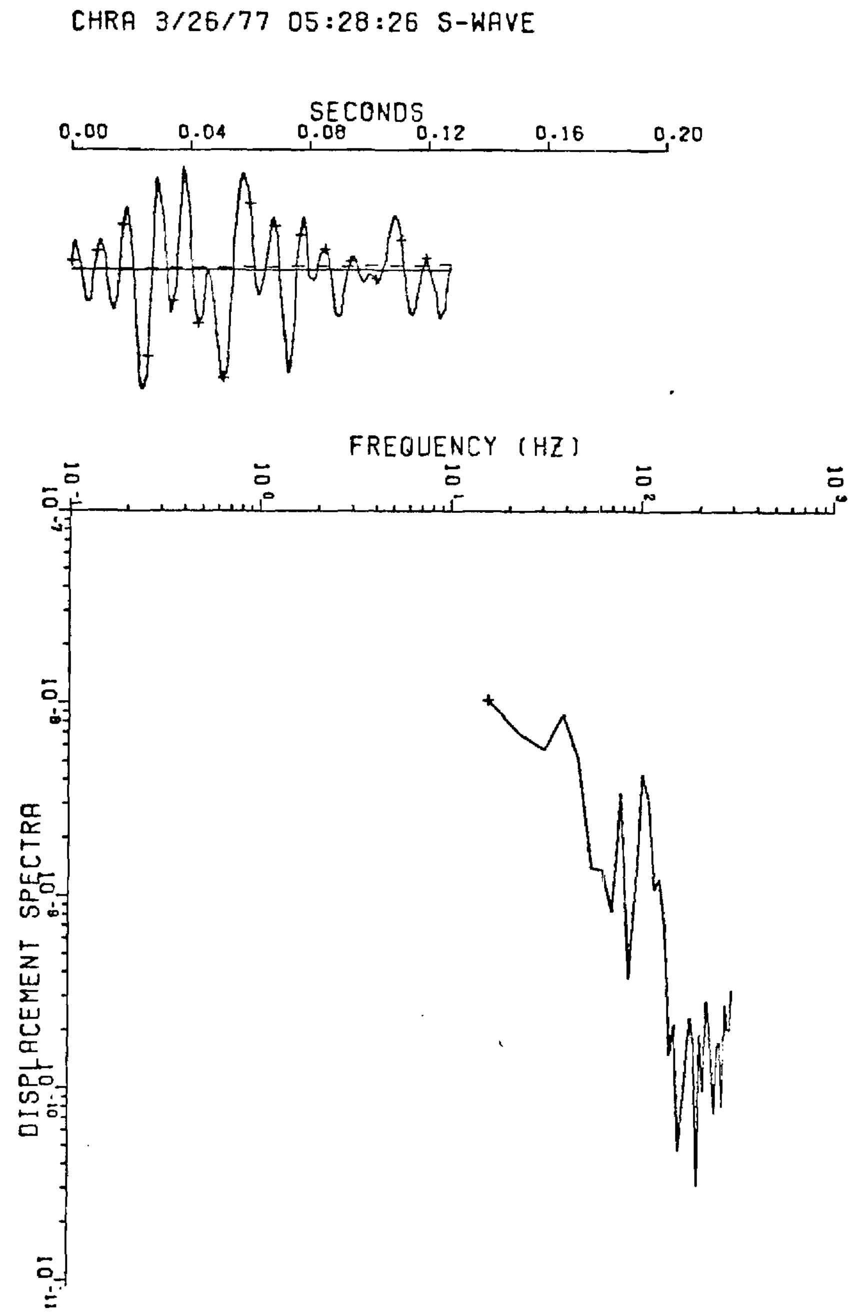

Figure 9. A typical Clark Hill Reservoir microearthquake displacement spectrum.

Table 3. Spectral Data For CHRA

ID	DATE	TIME	PHASE	SLOPE	RANGE	OMEGA	FREQ	RADIUS	MOMENT	STRESS	ML
1 DKF	770326	20403	P	2.8	1.20	.60E-08	95.	.017	.52E+16	.474	
2 DKF	770326	34957	P	3.0	1.10	.10E-08	85.	.019	.79E+15	.052	
2 DKF	770326	34957	S	3.3	1.10	.21E-08	90.	.013	.32E+15	.059	
3 DKF	770326	52826	P	3.0	1.10	.18E-08	110.	.015	.14E+16	.202	
3 DKF	770326	52826	S	3.5	1.10	.18E-08	105.	.011	.27E+15	.079	
4 DKF	770326	54844	P	2.9	.90	.80E-08	90.	.018	.52E+16	.403	
4 DKF	770326	54844	S	2.5	.90	.15E-07	65.	.018	.18E+16	.128	
5 DKF	770326	65056	P	3.0	1.30	.60E-08	60.	.027	.56E+16	.129	
5 DKF	770326	65056	S	3.0	1.30	.91E-08	50.	.024	.16E+16	.052	
6 DKF	770326	181352	P	3.0	1.40	.60E-08	90.	.018	.60E+16	.470	
7 DKF	770326	190443	P	3.2	1.20	.25E-08	110.	.015	.22E+16	.307	
7 DKF	770326	190443	S	4.0	1.20	.28E-08	100.	.012	.47E+15	.119	
8 DKF	770326	190458	P	2.5	1.90	.80E-08	70.	.023	.11E+17	.400	
9 DKF	770326	194306	P	3.0	.90	.25E-08	120.	.013	.16E+16	.299	
9 DKF	770326	194306	S	3.4	.90	.21E-08	110.	.011	.26E+15	.089	
10 DKF	770326	204242	P	3.0	.80	.14E-08	120.	.013	.81E+15	.149	
10 DKF	770326	204242	S	3.0	.80	.25E-08	100.	.012	.27E+15	.069	
11 DKF	770414	10530	P	2.0	1.60	.80E-09	60.	.020	.92E+15	.050	
11 DKF	770414	10530	S	2.5	1.60	.21E-08	55.	.022	.47E+15	.020	
12 DKF	770414	162242	P	3.5	1.10	.60E-09	140.	.011	.47E+15	.139	
12 DKF	770414	162242	S	2.5	1.10	.21E-08	90.	.013	.32E+15	.059	
13 DKF	770414	164540	P	3.0	1.10	.45E-08	70.	.023	.36E+16	.130	
13 DKF	770414	164540	S	3.0	1.10	.63E-08	70.	.017	.97E+15	.084	
14 DKF	770414	164717	P	3.5	1.50	.10E-08	140.	.011	.11E+16	.316	
14 DKF	770414	164717	S	3.0	1.50	.42E-08	80.	.015	.88E+15	.114	
15 CH6	760924	190500	P	3.0	12.10	0.	13.	.123	0.	0.000	1.8
16 CH6	760929	94814	P	3.0	7.60	0.	19.	.084	0.	0.000	1.4
17 CH5	780612	63300	P	2.8	27.30	.30E-07	18.	.089	.59E+18	.367	1.7
17 CH5	780612	63300	S	3.6	27.30	.56E-07	20.	.060	.21E+18	.432	1.7

(10) shows station and epicenter locations. An event of magnitude 2.8, which was also felt locally, was recorded at an epicentral distance of 2.0 km on a 3-component magnetic tape system. Accelerogram records of a magntidue 2.7 MRA were obtained from U. S. Geological Survey open file report. All of these were Fourier analyzed and the spectral slopes calculated. Table (4) lists magnitude, corner frequency, and exponential decay rates of Monticello Reservoir events. Most showed a decay of ω^{-3} and a fp/fs > 1. Preliminary results of focal mechanism analysis of the data shows normal or strike-slip faulting.

Lake Sinclair - Wallace Dam Area. Lake Sinclair is located in north central Georgia in the Piedmont Province about 100 km east of Atlanta. Natural seismicity has been reported in this area with a magnitude 4.0 earthquake being recorded in 1963. The Wallace Dam area 15 km northeast of Lake Sinclair has been monitored for seismic activity since March, 1977 by Georgia Tech. Some seismic activity has been spatially associated with the Lake Sinclair reservoir area. No studies have been possible to show whether this activity is triggered by the reservoir. Earthquake recordings on magnetic tape of the Lake Sinclair area earthquakes were examined to see if this data could be applicable for spectral analysis. Several earthquakes were determined to fit the criteria established for successful high frequency analysis as determined by magnitude epicentral distance, and attenuation levels. Fig. (11) shows station and epicenter location for these Lake Sinclair area events. On November 7, 1979 a magnitude 2.1 earthquake occurred near Lake Sinclair. A seismic network of smoked paper and magnetic tape recorders were quickly established near the estimated epicenter. One aftershock, which was the first microearthquake at Lake Sinclair to be

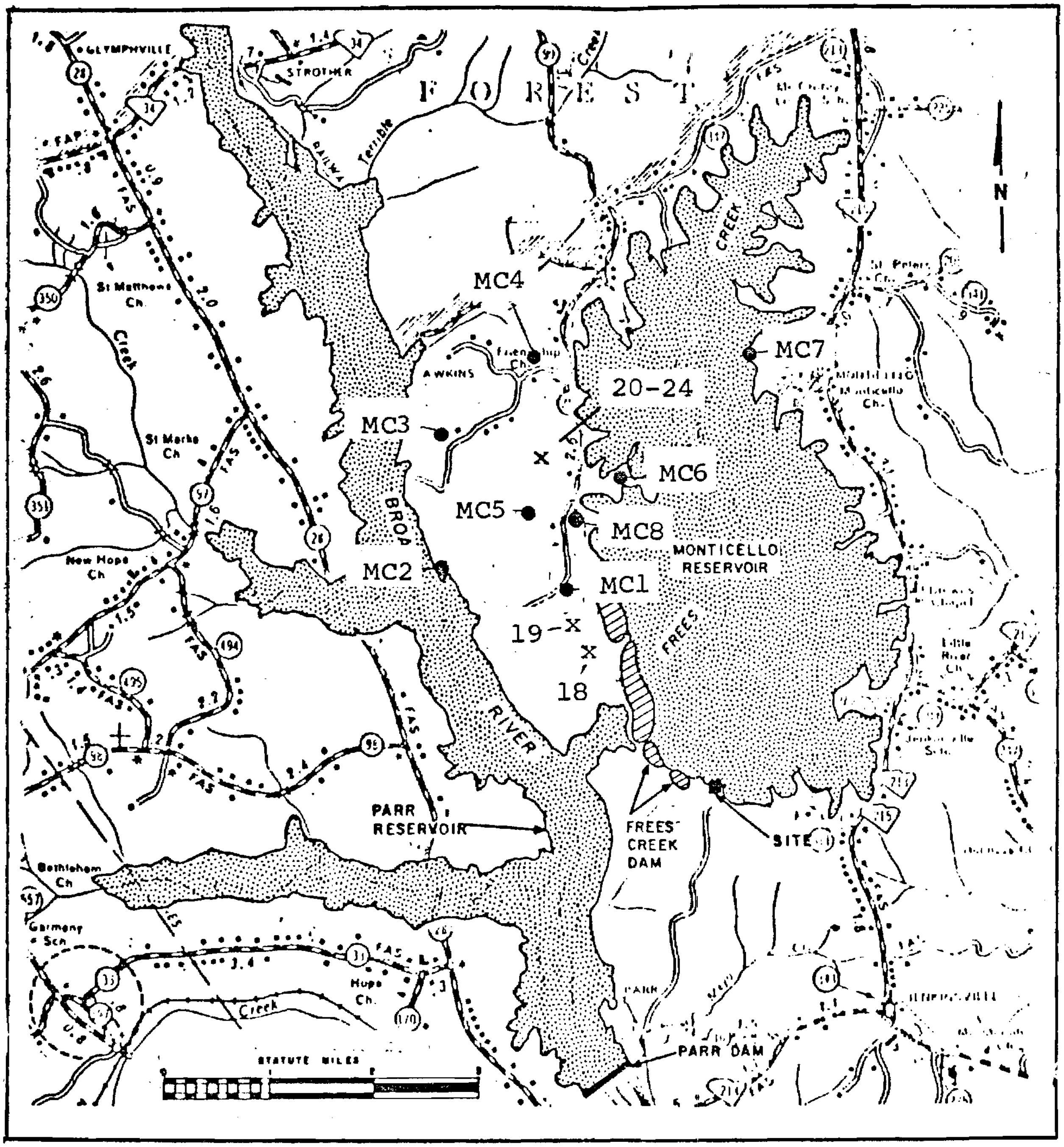

Figure 10. Monticello Reservoir temporary seismic station and earthquake locations.

Table 4. Spectral Data For MRA

ID	DATE	TIME	PHASE	SLOPE	RANGE	OMEGA	FREQ	RADIUS	MOMENT	STRESS	ML
18	USGS780827	102300	S	2.8	1.60	0.	8.	.150	0.	0.000	2.7
18	USGS780827	102300	S	3.4	1.60	0.	12.	.100	0.	0.000	2.7
18	USGS780827	102300	S	2.9	1.60	0.	12.	.100	0.	0.000	2.7
19	MC8 790806	194200	C	2.6	2.20	0.	7.	.171	0.	0.000	2.8
19	MC8 790806	194200	C	2.8	2.20	0.	6.	.188	0.	0.000	2.8
19	MC8 790806	192400	C	2.7	2.20	0.	6.	.194	0.	0.000	2.8
20	MC7 790615	224200	P	3.0	3.20	0.	90.	.018	0.	0.000	
20	MC7 790615	224200	C	3.0	3.20	0.	105.	.011	0.	0.000	
20	MC7 790615	224200	S	2.5	3.20	0.	65.	.018	0.	0.000	
21	MC4 791116	142922	P	3.1	1.50	.20E-09	60.	.027	.22E+15	.005	
21	MC4 791116	142922	S	3.2	1.50	.63E-09	45.	.027	.13E+15	.003	
22	MC4 791116	151536	P	3.0	1.70	.30E-08	45.	.036	.37E+16	.036	
22	MC4 791116	151536	S	3.0	1.70	.35E-08	50.	.024	.83E+15	.026	
23	MC4 791116	152421	P	3.0	1.90	.15E-08	45.	.036	.20E+16	.020	
23	MC4 791116	152421	S	2.9	1.90	.32E-08	40.	.030	.83E+15	.014	
24	MC4 791116	152437	P	3.0	1.90	.10E-07	40.	.040	.14E+17	.093	
24	MC4 791116	152437	S	3.2	1.90	.28E-07	30.	.040	.74E+16	.051	

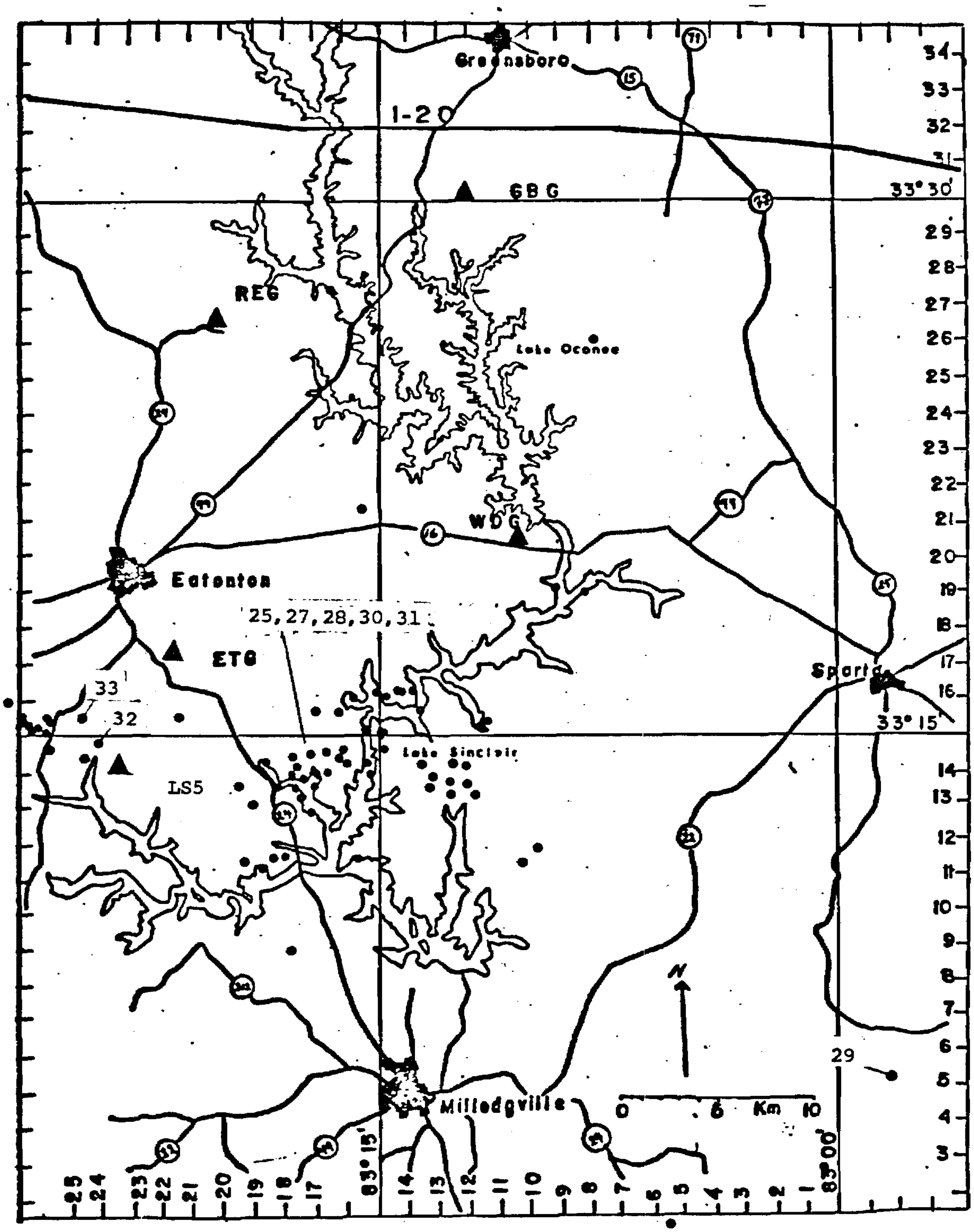

Figure 11. Location map for Lake Sinclair - Wallace Dam seismic activity and stations.

recorded on a high frequency magnetic tape system, has been frequency analyzed. Fig. (12) shows the microearthquake P-wave spectra obtained. Table (5) lists magnitudes, corner frequencies and spectral decay of Lake Sinclair area events. Most of the Lake Sinclair spectra typically decay as ω^{-2} and have fp/fs $\leq$ 1.0.

IX Q Calculation for the Georgia - South Carolina Piedmont Province

The amplitude of the high-frequency spectra can be severely attenuated at high-frequencies by absorptive attenuation which varies with raypath, depth, and crustal composition. The criteria which was developed for reliably computing spectra from telemetry systems depends on the Q value for defining the reliably observed high frequencies. Accordingly, Q was computed for the Piedmont Province in which all three of our data areas are located. The method of spectral ratios (Der and McElfresh, 1977) was used for earthquakes recorded locally and regionally to get a Q value for the near surface crust. Fig. (13) shows the spectral ratio of an event recorded by two stations at epicentral distances of 20 to 50 km. By this method, 6 ratios were determined (see Appendix I). A frequency independent Q for the upper crust in the Piedmont was found to be 760 $\pm$ 100 in the frequency range of 1.0 to 30 Hertz.

X. Conclusion

Calculated Fourier displacment spectra of earthquakes recorded at the Clark Hill, Jocassee, and Monticello reservoir areas show an ω^{-3} high-frequency spectral decay. Spectra from earthquakes of the Wallace Dam - Lake Sinclair area, which has not been shown to be a reservoir induced seismic area, show an ω^{-2} decay. No seismicity has been associated with the Wallace Dam area after one year of filling. It

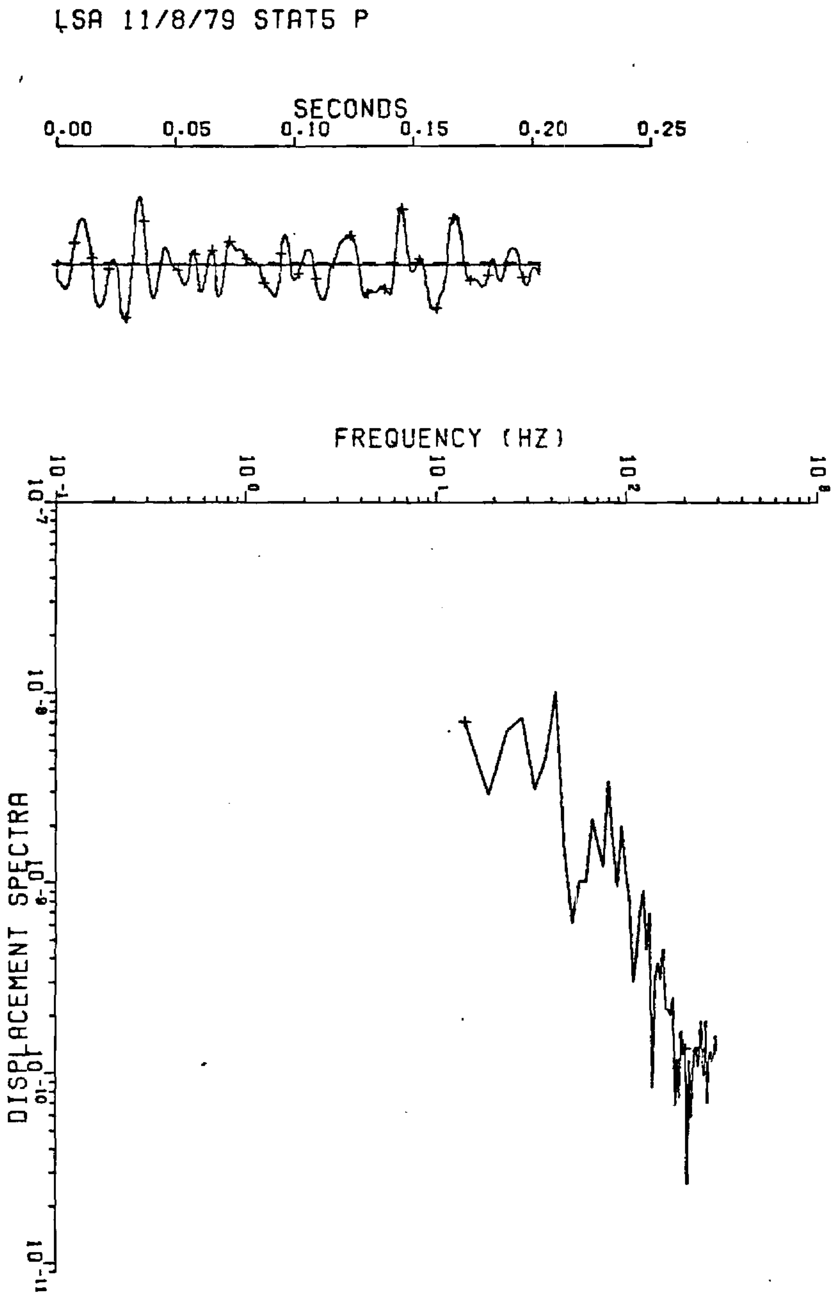

Figure 12. P-wave spectra of a Lake Sinclair area microearthquake.

Table 5. Spectral Data For Lake Sinclair-Wallace Dam Area

ID	DATE	TIME	PHASE	SLOPE	RANGE	OMEGA	FREQ	RADIUS	MOMENT	STRESS	ML
25 ETG	770831	165600	P	1.6	10.70	0.	12.	.100	0.	0.000	1.2
25 ETG	770831	165600	S	1.5	10.70	0.	20.	.040	0.	0.000	1.2
25 REG	770831	165600	P	1.5	24.20	0.	13.	.092	0.	0.000	1.2
26 REG	770906	200400	P	2.0	44.00	0.	10.	.120	0.	0.000	1.9
26 GBG	770906	200400	P	2.5	45.00	0.	15.	.107	0.	0.000	1.9
26 GBG	770906	200400	S	3.0	45.00	0.	18.	.067	0.	0.000	1.9
27 GBG	771123	223000	P	1.5	30.80	0.	16.	.075	0.	0.000	1.6
27 GBG	771123	223000	S	2.1	30.80	0.	16.	.075	0.	0.000	1.6
27 ETG	771123	223000	P	2.3	7.20	0.	20.	.080	0.	0.000	1.6
28 GBG	771125	90400	P	2.0	27.20	0.	10.	.120	0.	0.000	2.1
28 GBG	771125	90400	S	2.5	27.20	0.	14.	.086	0.	0.000	2.1
29 WDG	780320	122600	P	2.9	32.00	0.	12.	.133	0.	0.000	1.8
29 WDG	780320	122600	S	2.5	32.00	0.	16.	.075	0.	0.000	1.8
30 REG	780501	213000	P	2.0	23.80	0.	20.	.060	0.	0.000	1.6
30 REG	780501	213000	S	0.0	23.80	0.	23.	.035	0.	0.000	1.6
31 WDG	780502	14600	P	1.7	17.30	0.	9.	.133	0.	0.000	2.3
31 GBG	780502	14600	P	2.3	30.00	0.	11.	.145	0.	0.000	2.3
32 GBG	781002	2500	P	1.8	33.20	0.	16.	.075	0.	0.000	2.1
32 WDG	781002	2500	P	2.1	27.10	0.	20.	.080	0.	0.000	2.1
33 LS5	791108	235600	P	2.5	4.20	0.	80.	.020	0.	0.000	
33 LS5	791108	235600	S	2.3	4.20	0.	75.	.016	0.	0.000	

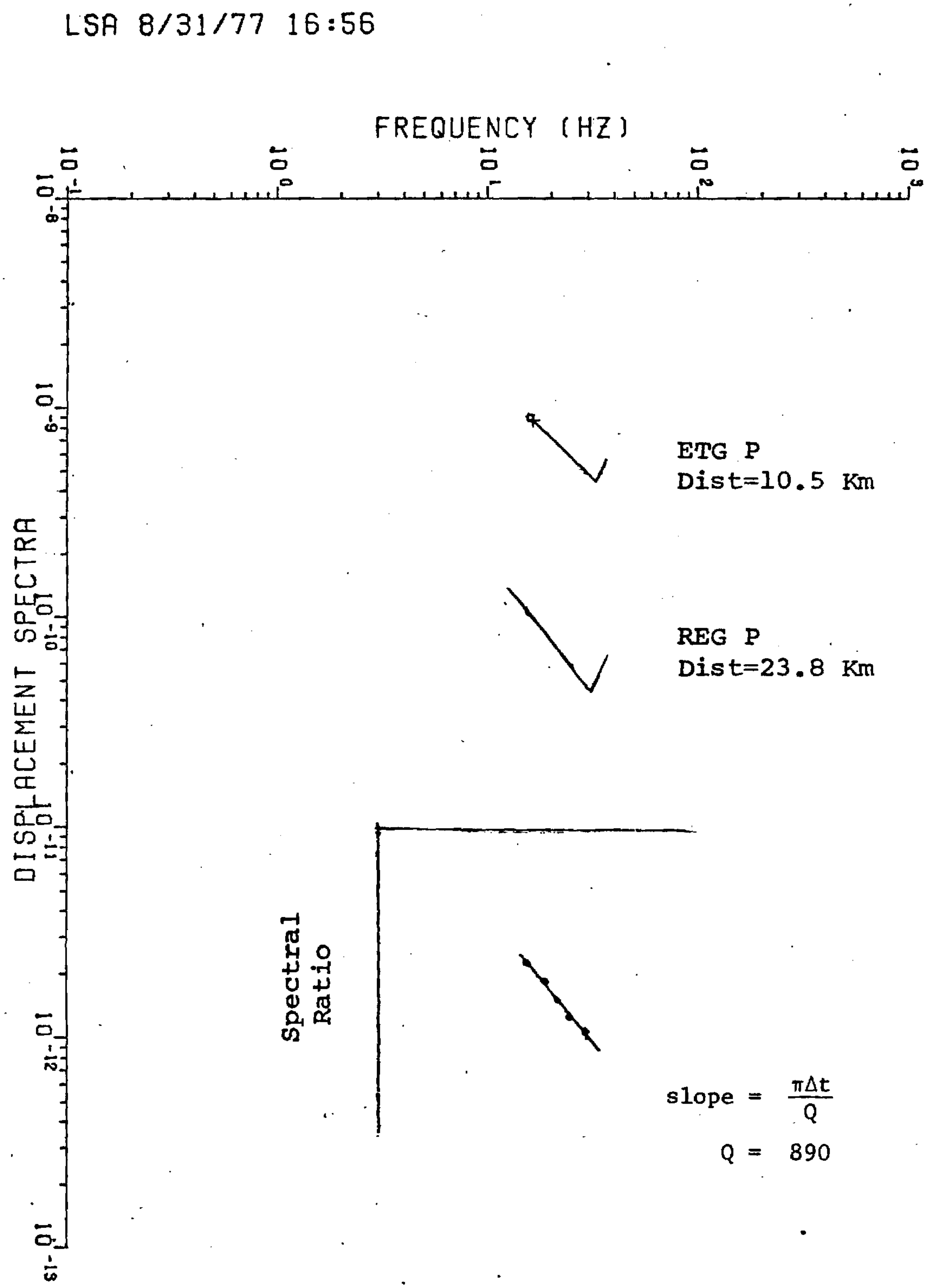

Figure 13. Spectral ratios as used to determine Q values for the Southeastern Piedmont Province.

appears that in the Georgia - South Carolina Piedmont province an ω^{-3} slope is indicative of areas where reservoirs have triggered seismic activity.

The sparsity of viable data from other reservoir areas limits the general evaluation of the seismic spectral discriminate. However, spectra of aftershocks of the Oroville, California magnitude 5.7 earthquake of August 1, 1975, considered to be reservoir associated, by some, have shown an ω^{-3} or greater decay (Fletcher, 1979) in an area of normal or strike-slip faulting (Bufe et al., 1976; Ryall et al., 1976). Studies of focal mechanisms and source parameters at Lake Mead were made by Rogers and Gallanthine (1973) and Rogers and Lee (1976). A right-lateral strike slip motion on vertical fault planes was generally observed. For some events, normal faulting was possible. Source parameters obtained for a group of seven Lake Mead events (S-waves) showed a high frequency decay of $\omega^{-1.3}$ to $\omega^{-1.8}$. This finding does not appear to support the spectral discriminant. However, the highest level of seismic activity at Lake Mead occurred not long after impounding in 1939. The length of time from impounding to the recording of recent data is perhaps sufficiently long that the stress field of the reservoir area has reached equilibrium and earthquake are no longer being induced by the reservoir. Rogers and Gallanthine (1973) conclude that tectonic forces appear to be chiefly responsible for the current seismicity at Lake Mead. With the exception of Lake Mead, available evidence supports an ω^{-3} decay in reservoir areas which have reportedly induced seismic activity and for which sufficient data are available. Reservoirs showing induced seismic activity are typically characterized by normal or strike-slip faulting. Of the major cases of reservoir induced seismicity, strike-slip or normal fault plane solutions were observed (Simpson, 1976). Reservoir effects on a pre-existing stress field can enhance rupture only under specific conditions. Increased vertical stress due to reservoir loading and decreased effective strength due to increased pore pressure can modify the stress regime in a reservoir area. These stress changes can initiate earthquakes dependent on the geologic and hydrologic conditions near the reservoir site. Regions where the deviatoric stess conditions are tensional have the greatest tendency for induced seismic activity. In regions where thrust faulting dominates, stresses due to a vertical load should have a minimum effect on local seismicity (Simpson, 1976; Jacob et al., 1979).

Theoretical models of the seismic source support the use of the spectral slope to discriminate areas with compressive deviatoric stress from areas with tensional deviatoric stress. In general, the theoretical models indicate that subsonic rupture velocities on a fault with irregular tractional resistance, which is related to the material properties and normal stress, give spectra with multiple corner frequencies and -square or less decay at high frequencies. Smooth or lubricated faults with tractional resistance less than about 5 times the driving shear stress allow transonic rupture velocities and generate spectra with - cube decay. The latter description is expected near reservoirs for shallow joints and faults believed to be involved in the reservoir induced seismic activity. However, the above proposed deviatoric stress conditions and spectral discriminant may be a function also of depth of focus since overburden pressure would increase the normal stress on a fault plane and hence increase its tractional resistance. All of the reservoir induced events we recorded were shallow (less than 2 km). However, the depths of the Sinclair area events were not found but are probably less than 3 km.

The Oroville data, however, cover a wide range of depths and are shown in Fig. 14. The data of Fig. 14 do not show a significant preference for lesser spectral decay with increased depth. Hence our limited existing data indicate that depth may not be a significant factor in the application of the discriminant.

XI Recommendations

1. Because the discriminant appears to work, we recommend it be applied in conjunction with other seismic information prior to the filling of a reservoir.

2. All new reservoirs should be monitored with seismic equipment capable of generating data which can be used to compute the high-frequency spectra.

3 Further studies should be performed on the variation of spectral high-frequency slope with increased depth of focus.

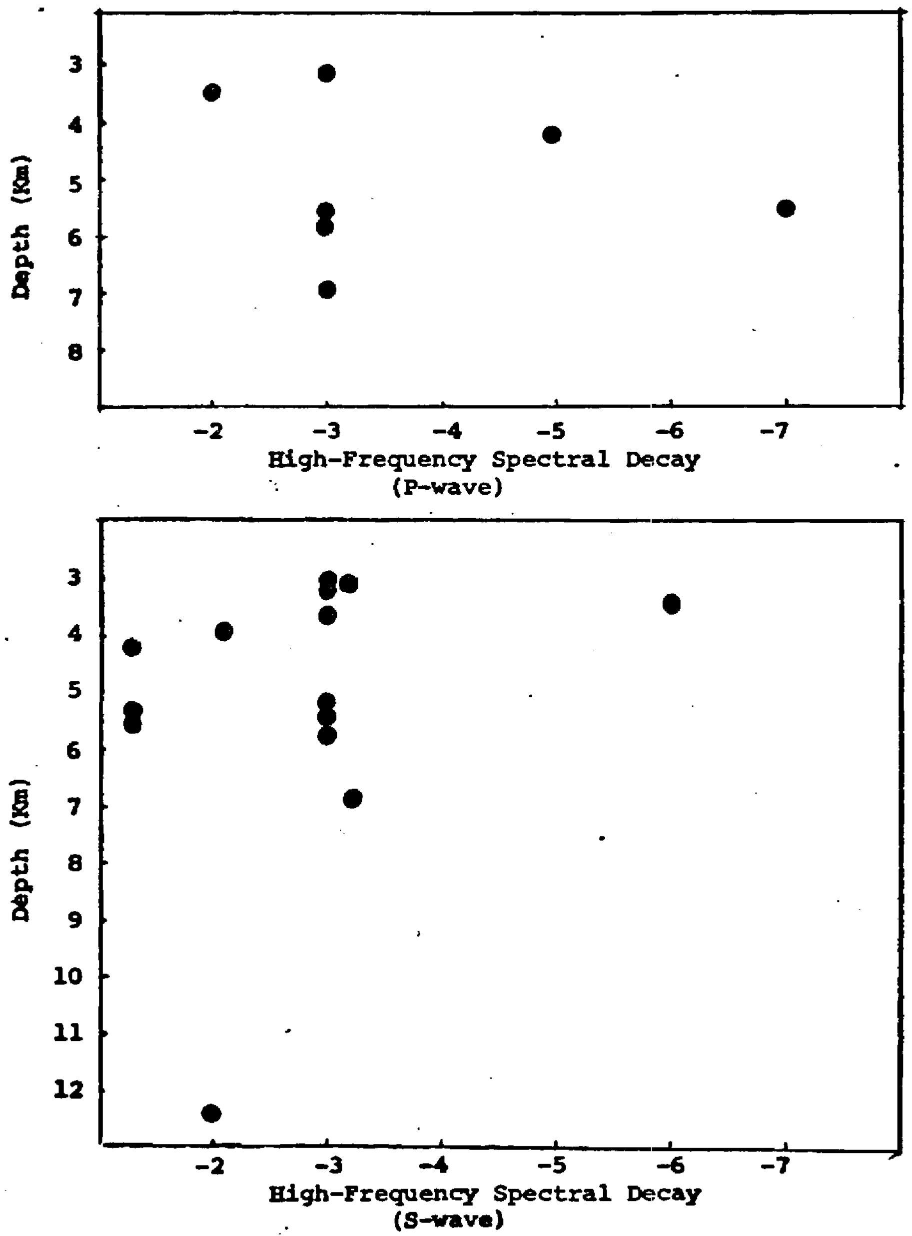

Figure 14. Plot of depth versus high-frequency spectral decay rate for Oroville aftershocks (after Fletcher, 1979).

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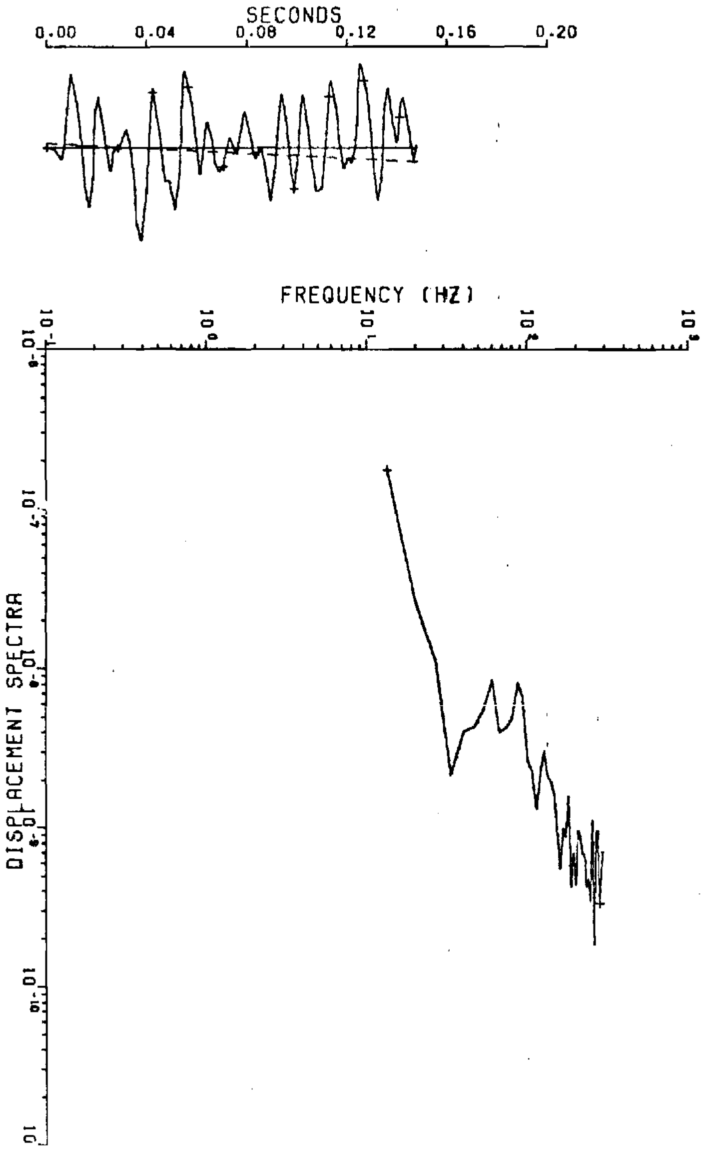

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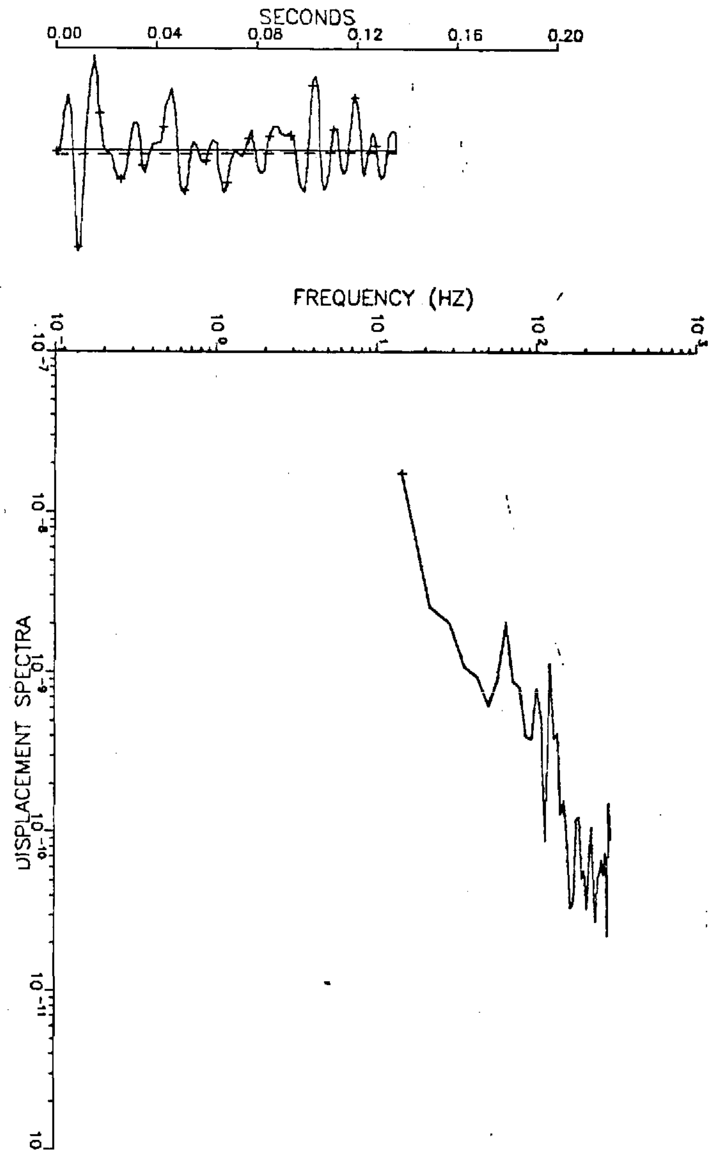

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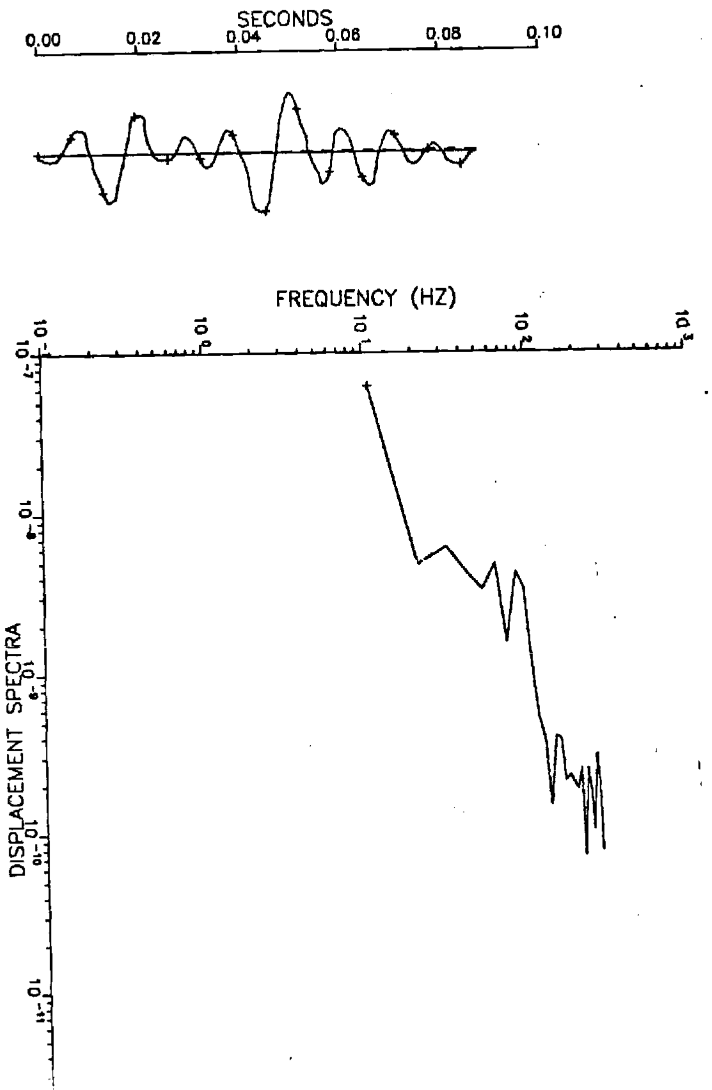

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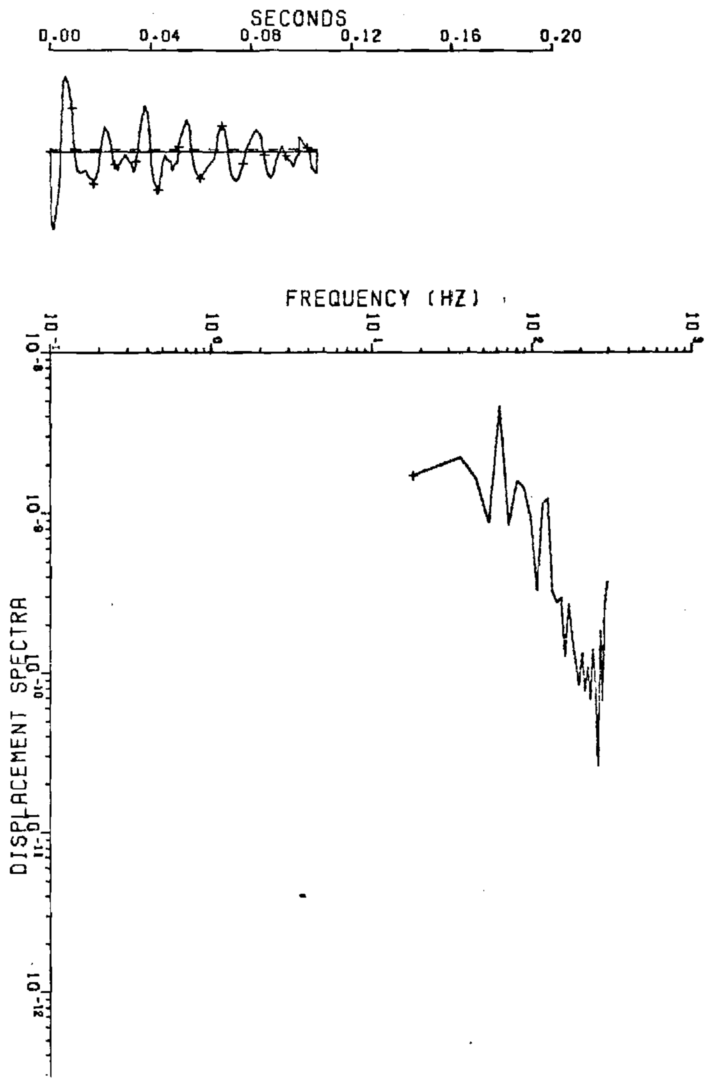

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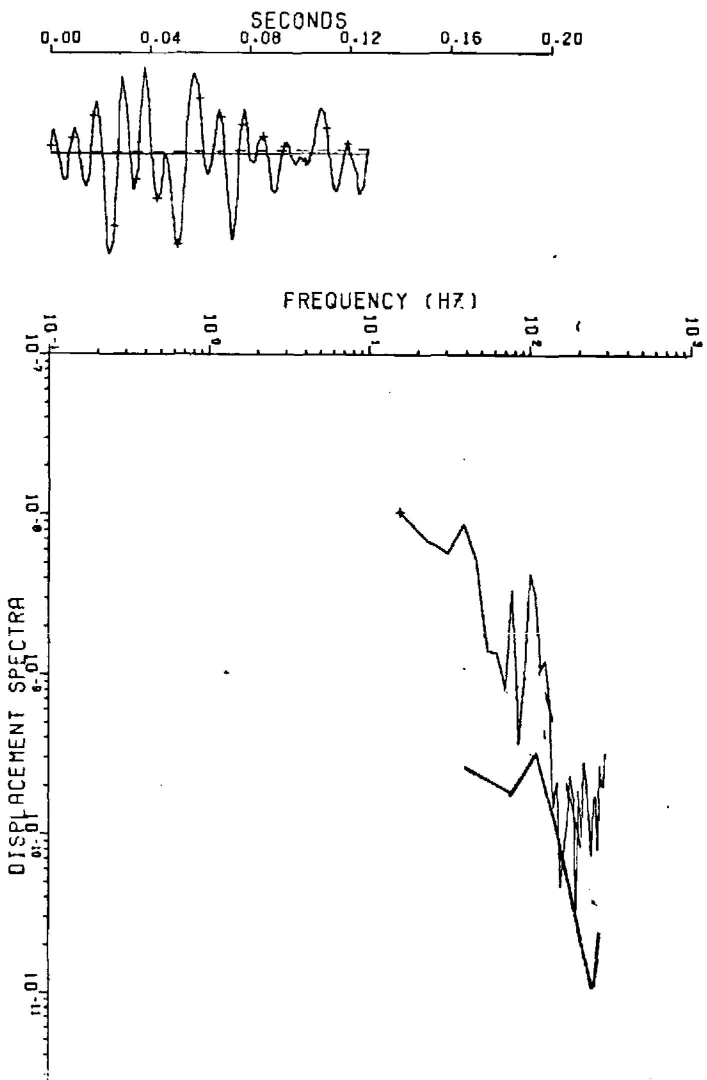

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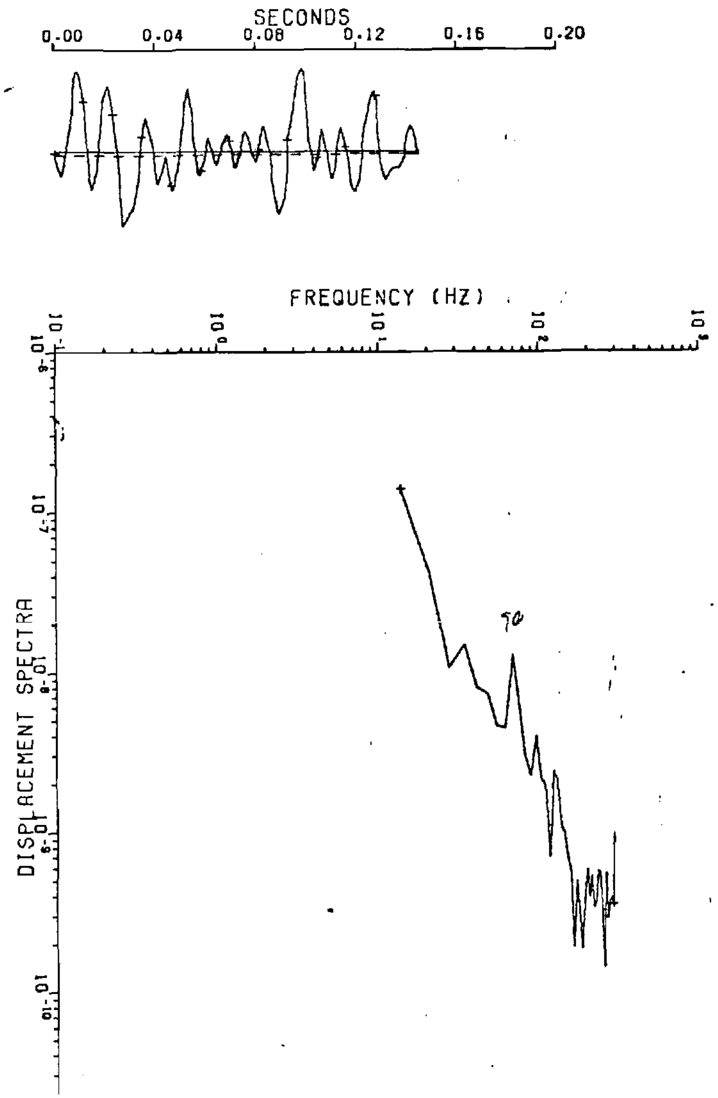

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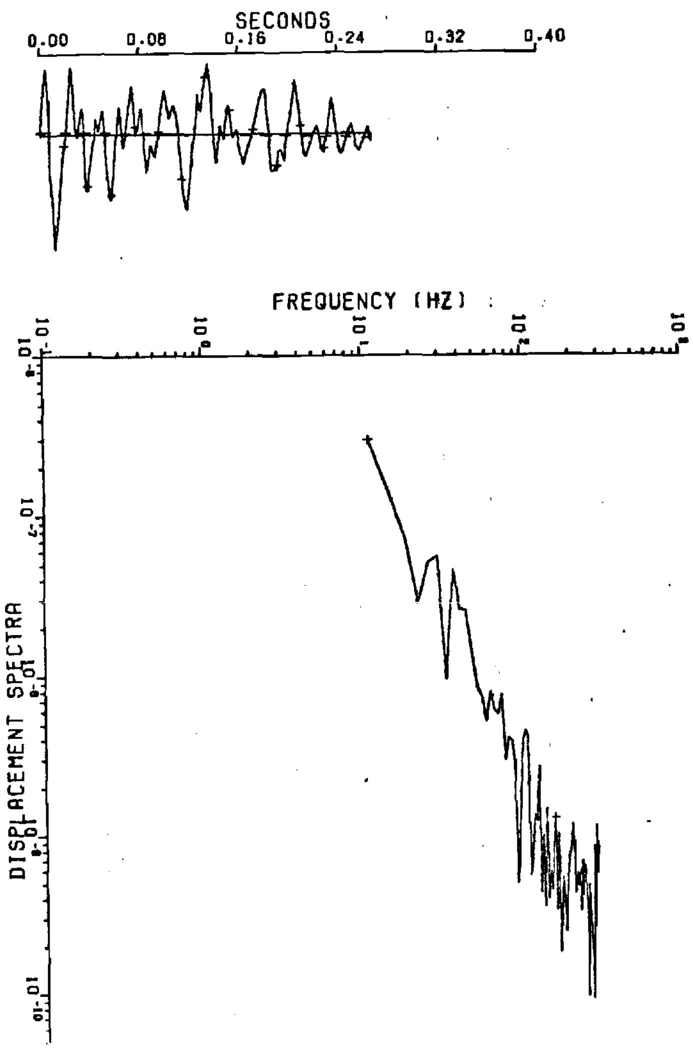
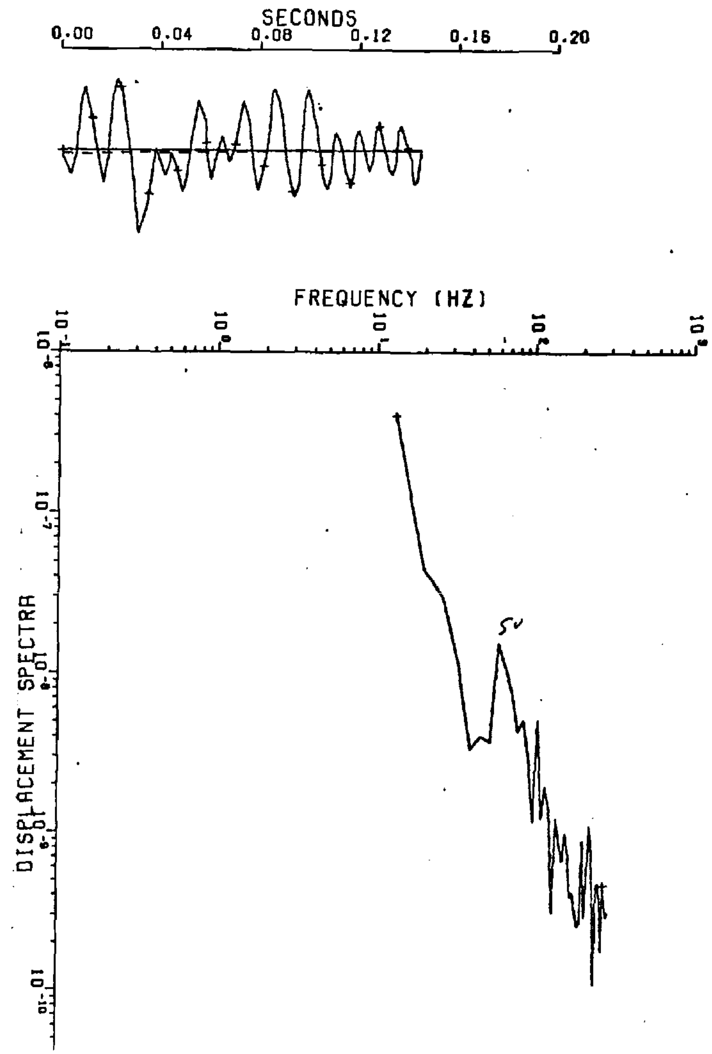
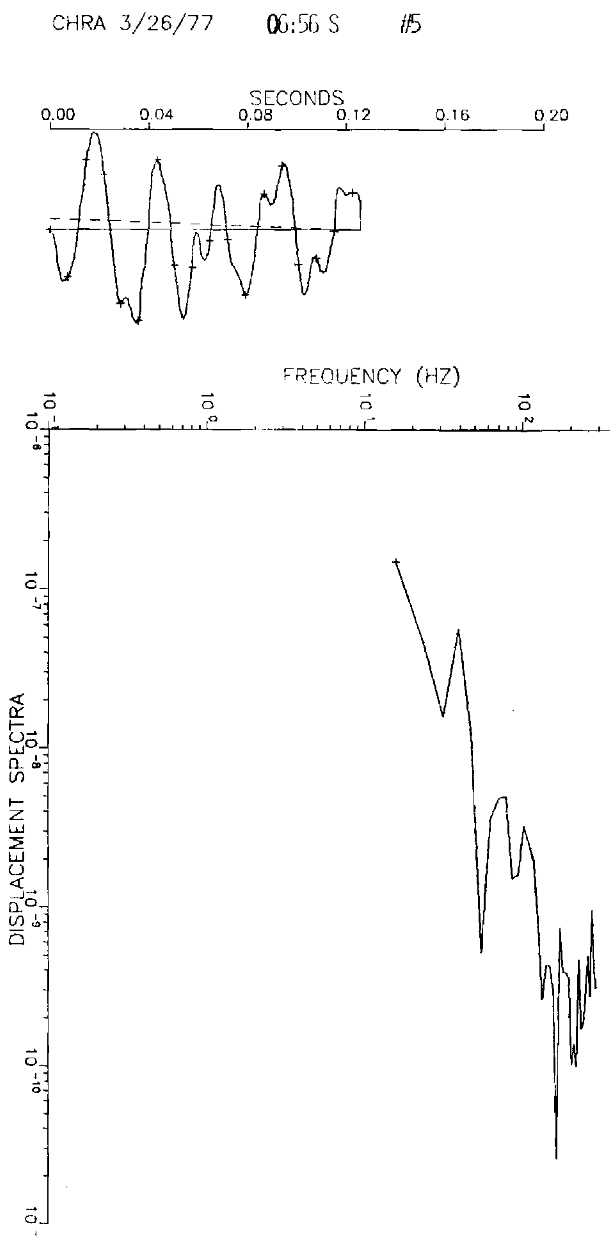
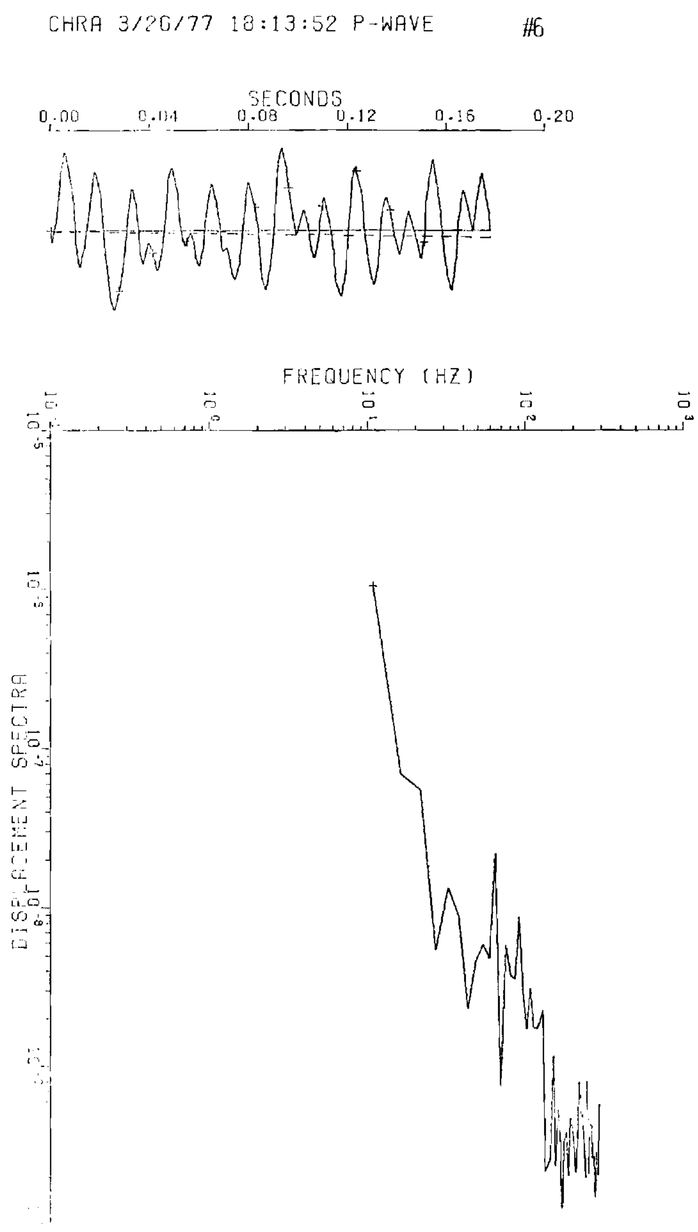


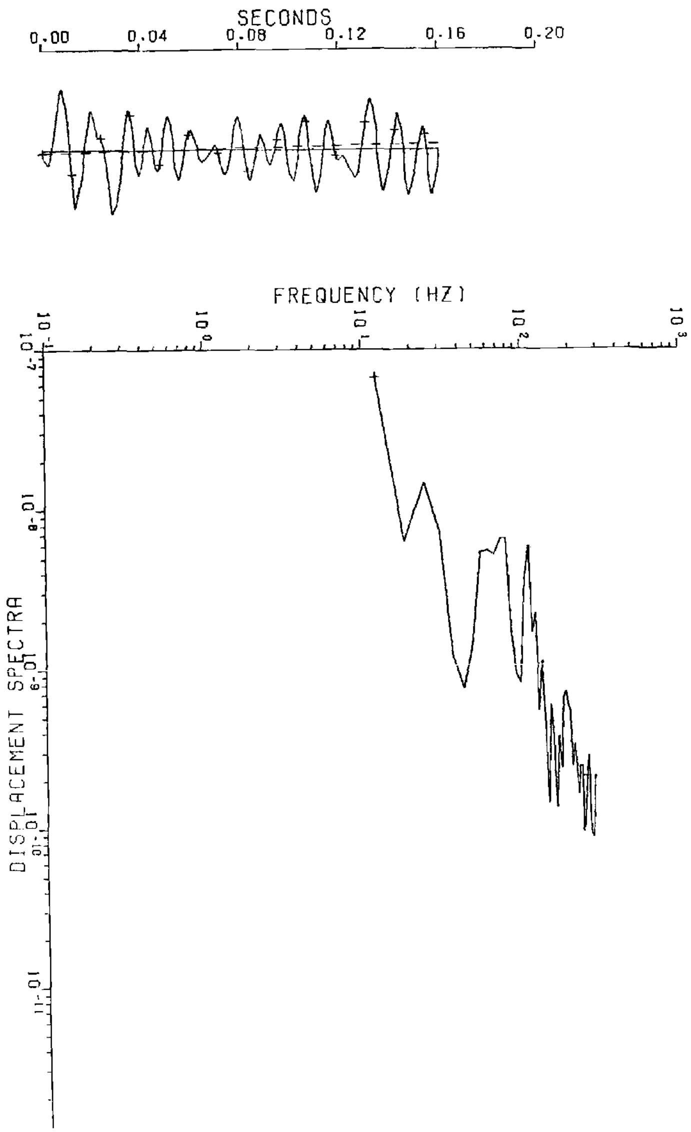
CHRA 3/26/77 06:50:56 P-WAVE #5



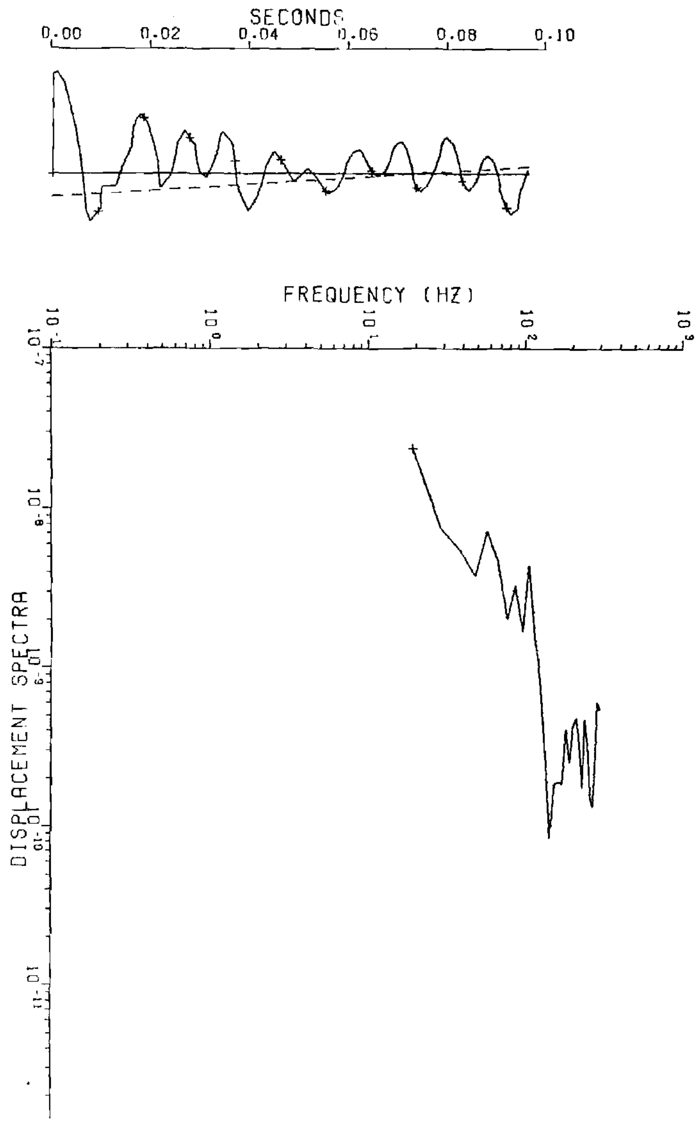

CHRA 3/26/77 06:56 S #5
SECONDS
0.00
0.04
0.08
0.12
0.16
0.20
FREQUENCY (HZ)
DISPLACEMENT SPECTRA


CHRA 3/26/77 18:13:52 P-WAVE #6
SECONDS
0.00
0.04
0.08
0.12
0.16
0.20
FREQUENCY (HZ)
DISPLACEMENT SPECTRA

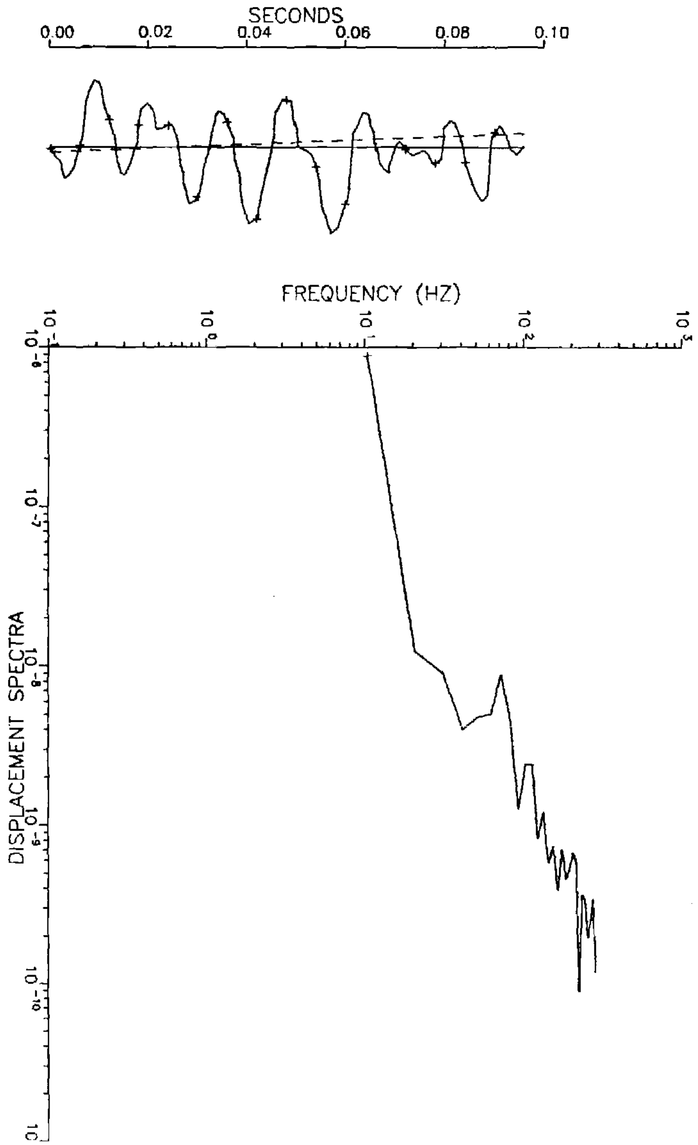
CHRA 3/26/77 19:04:43 P #7



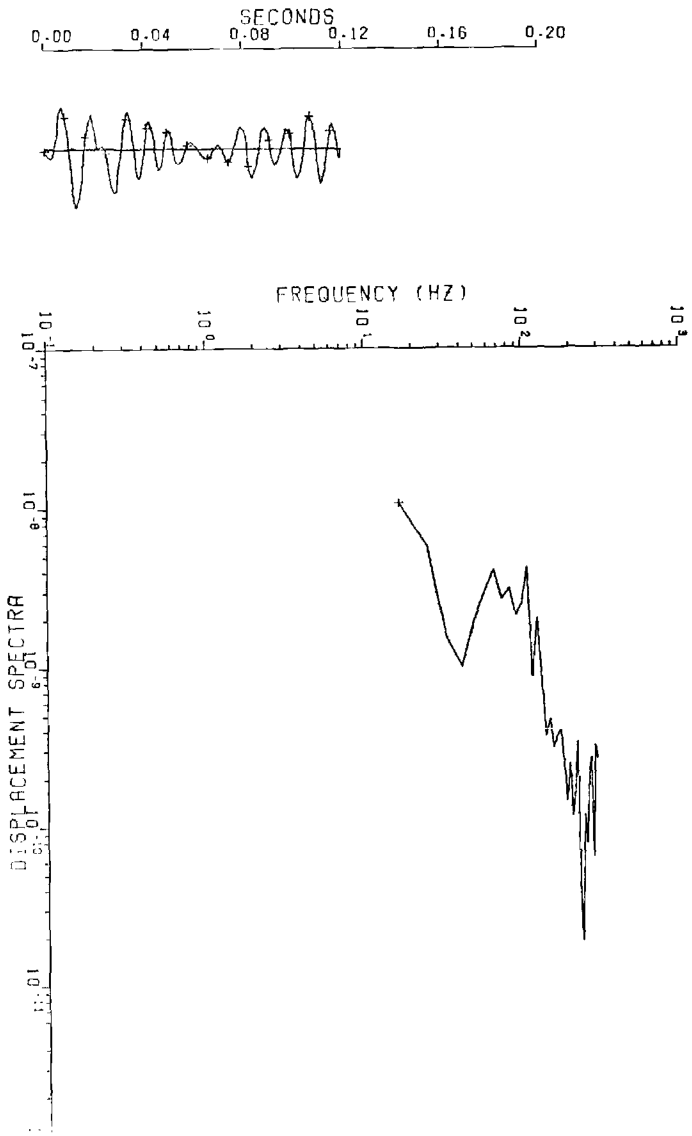
CHRA 3/26/77 19:04:43 S-WAVE #7



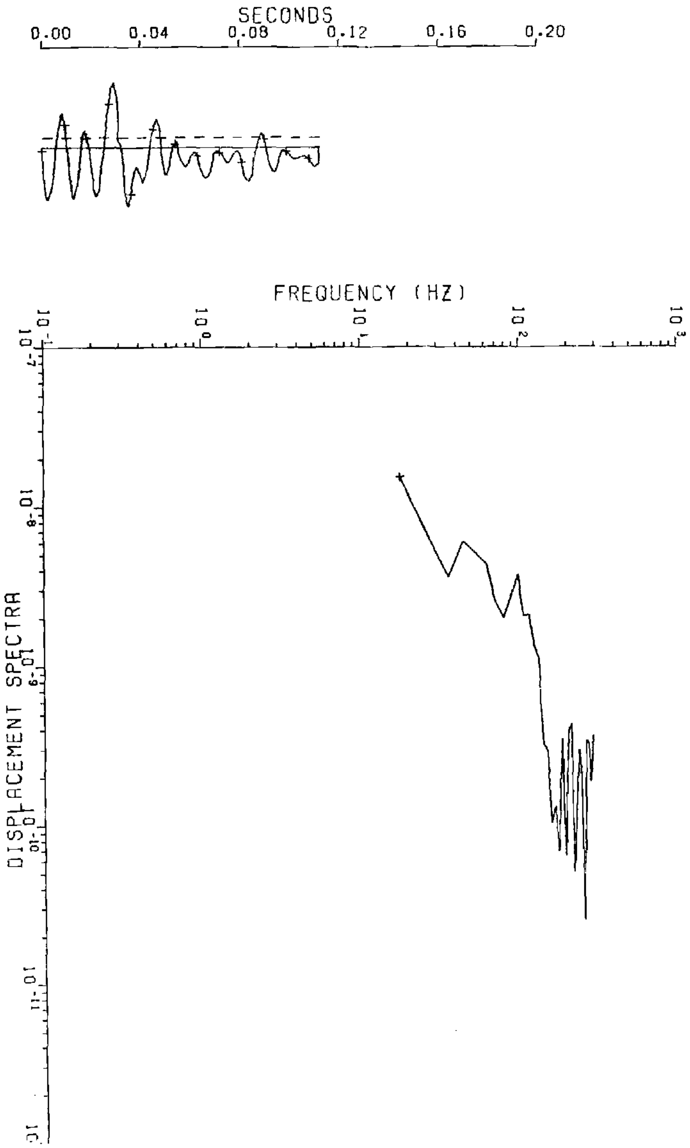
CHRA 3/26/77 19:04:58 P #8



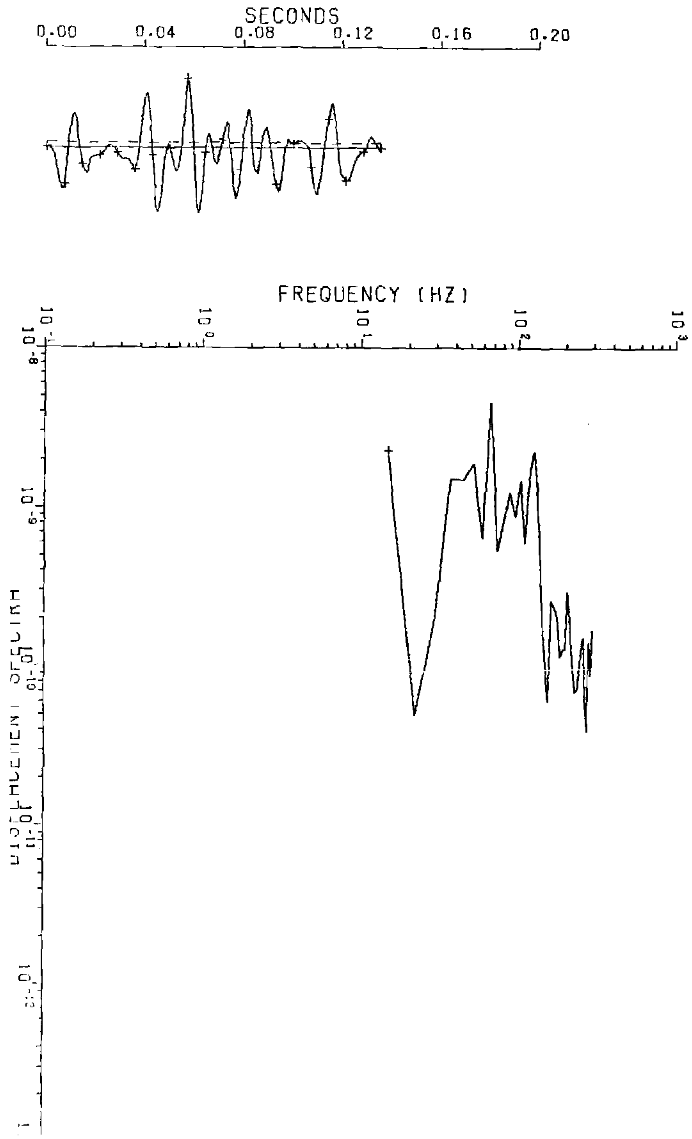
CHRA 3/26/77 19:43:06 P-WAVE #9



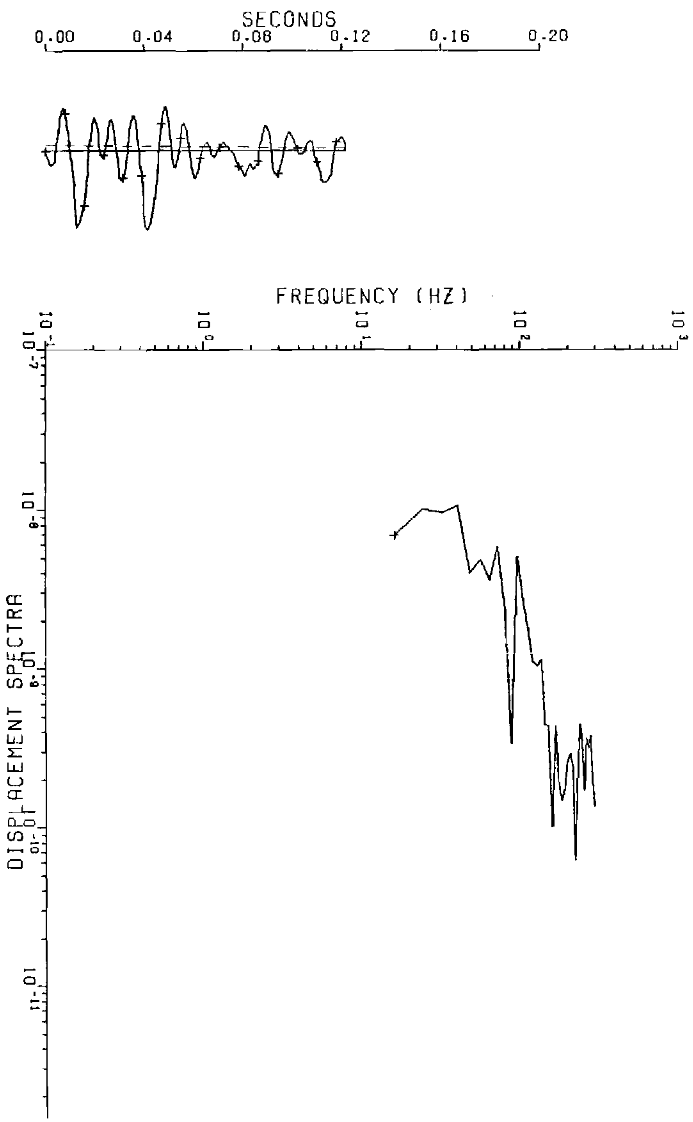
CHRA 3/26/77 19:43:06 S-WAVE #9



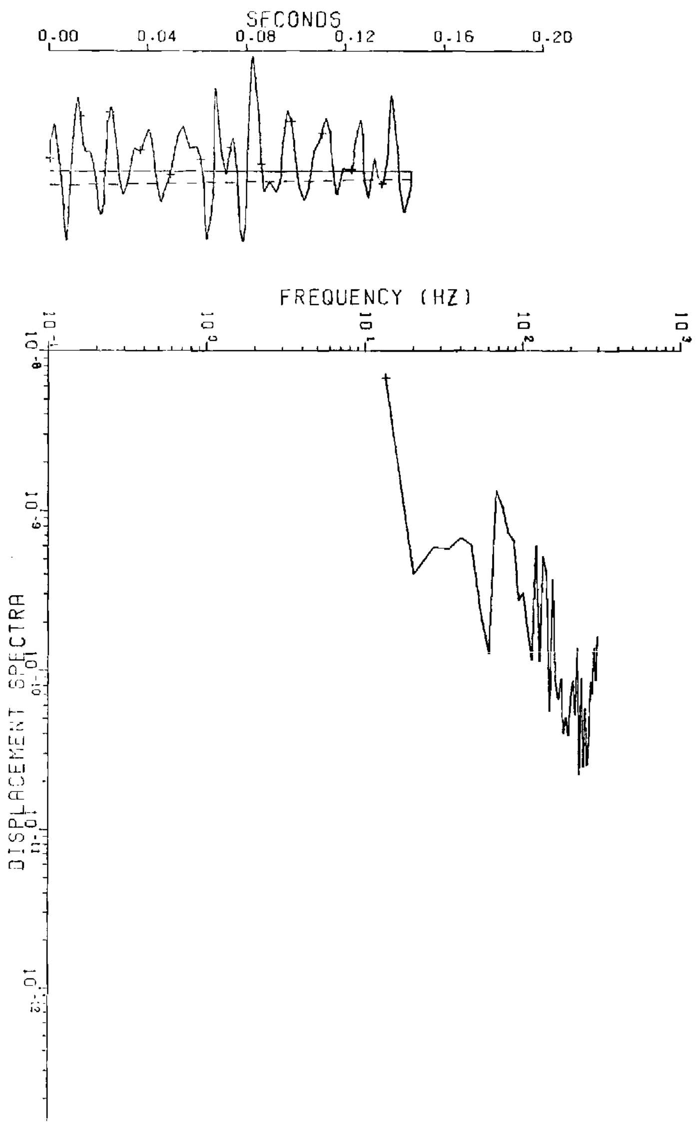
CHRA 3/26/77 20:42:42 P-WAVE #10



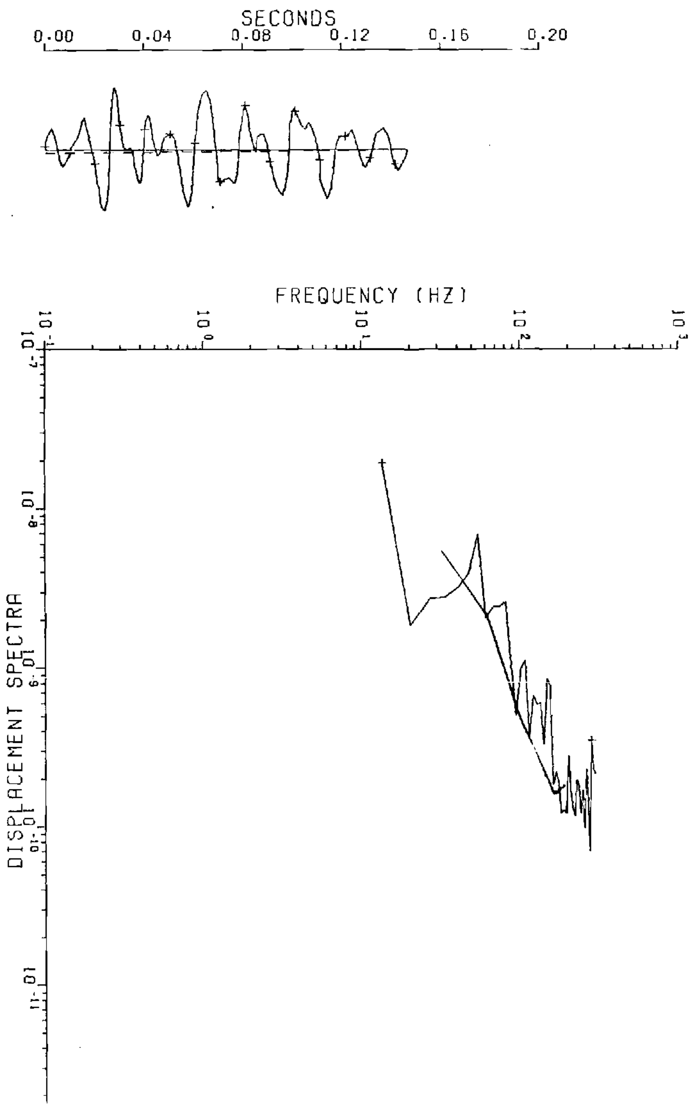
CHRA 3/26/77 20:42:42 S-WAVE #10



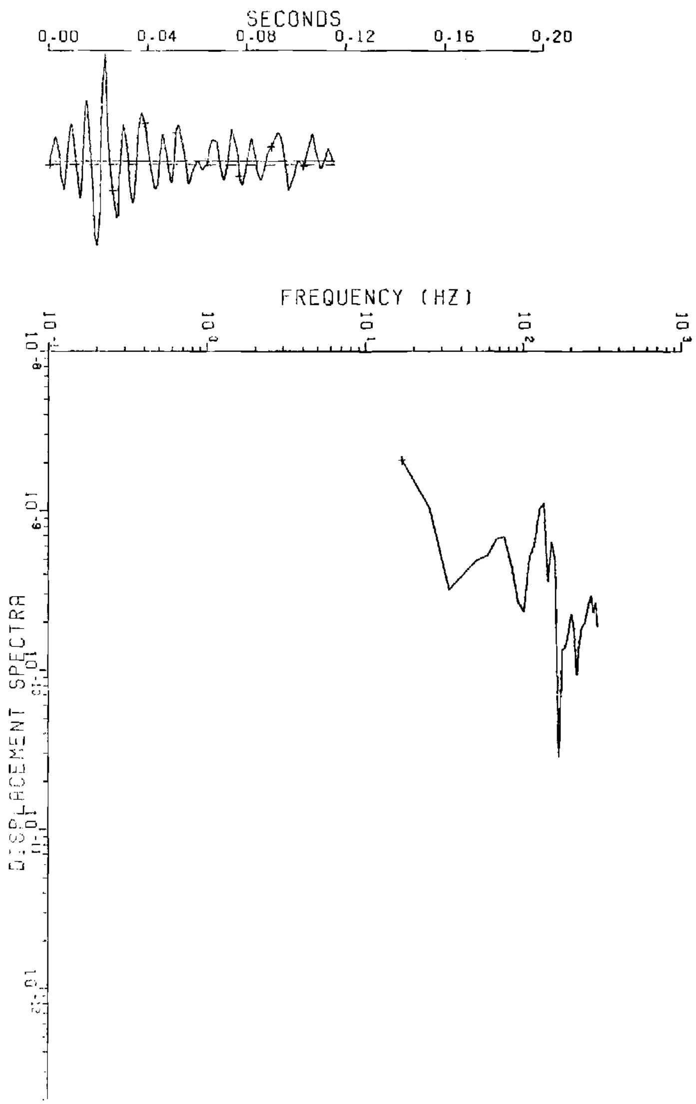
CHRA 4/14/77 01:05:30 P-WAVE #11



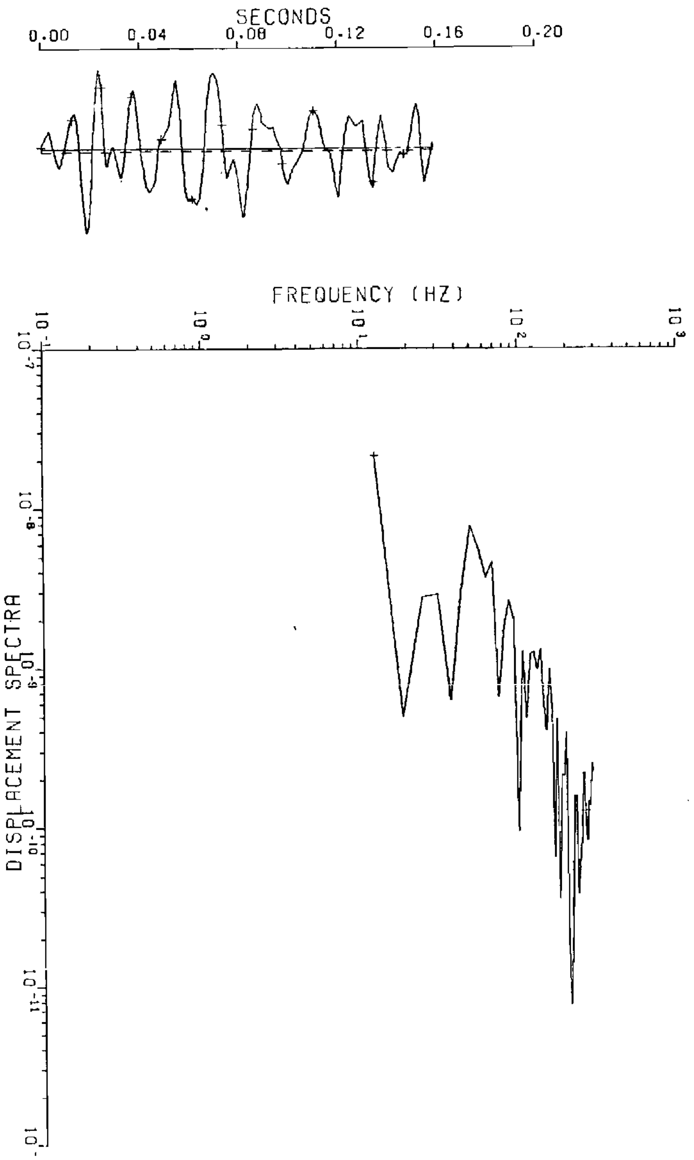
CHRA 4/14/77 01:05:30 S-WAVE #11



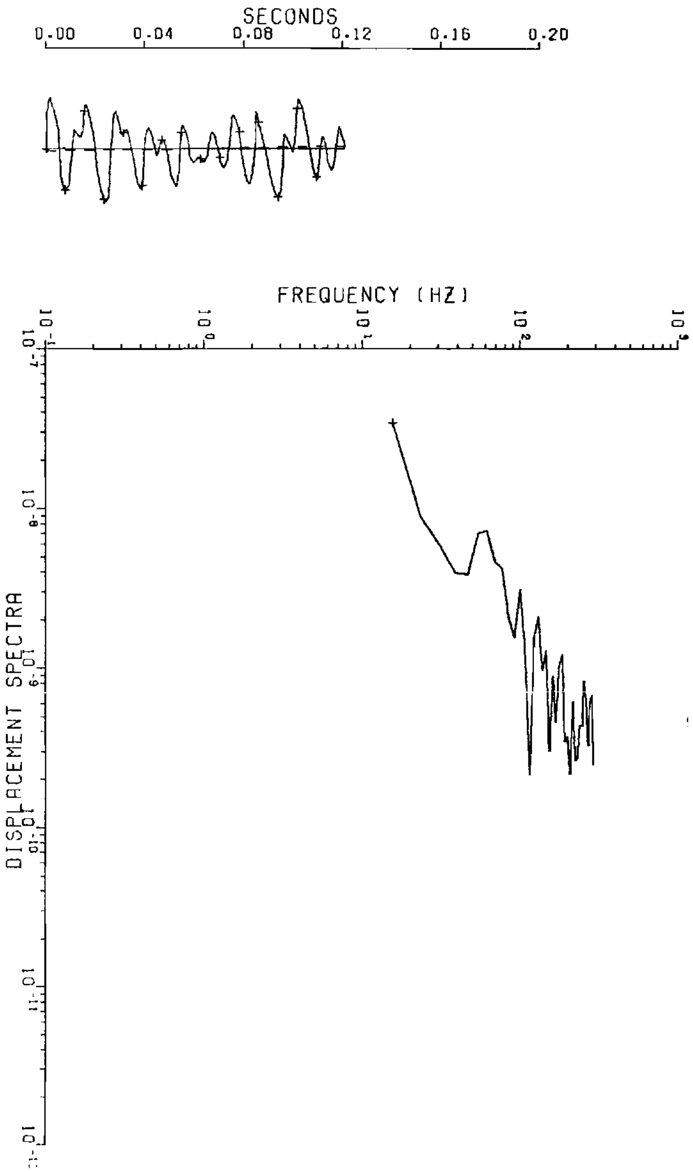
CHRA 4/14/77 16:22:42 P-WAVE #12



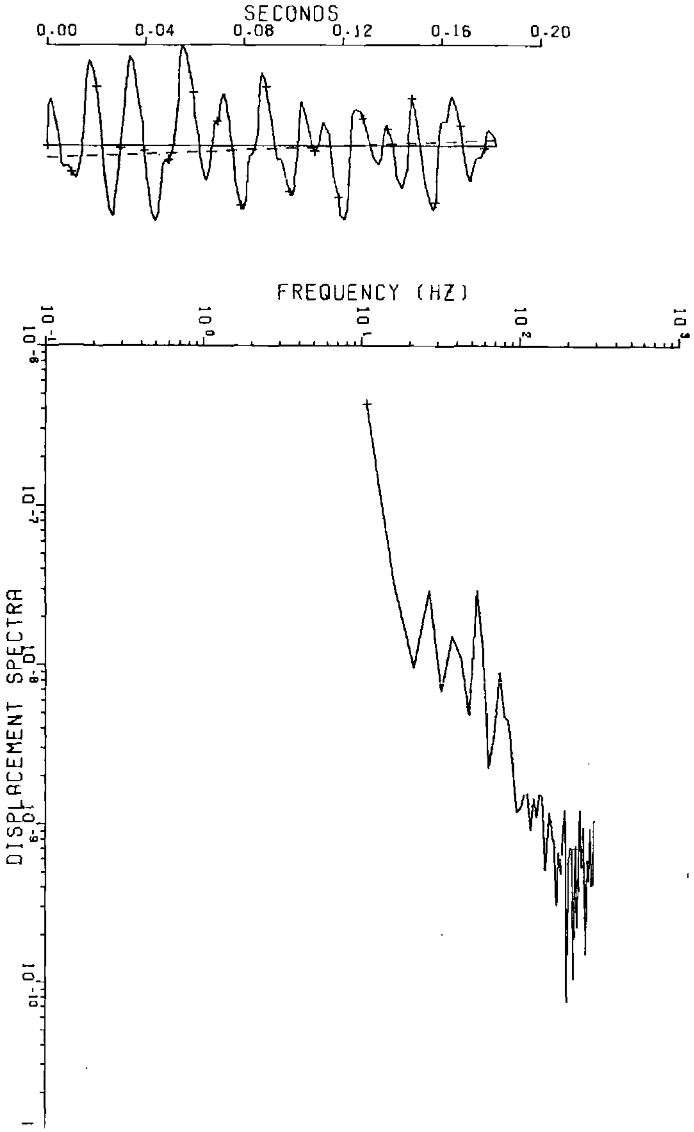
CHRA 4/14/77 16:22:42 S-WAVE #12



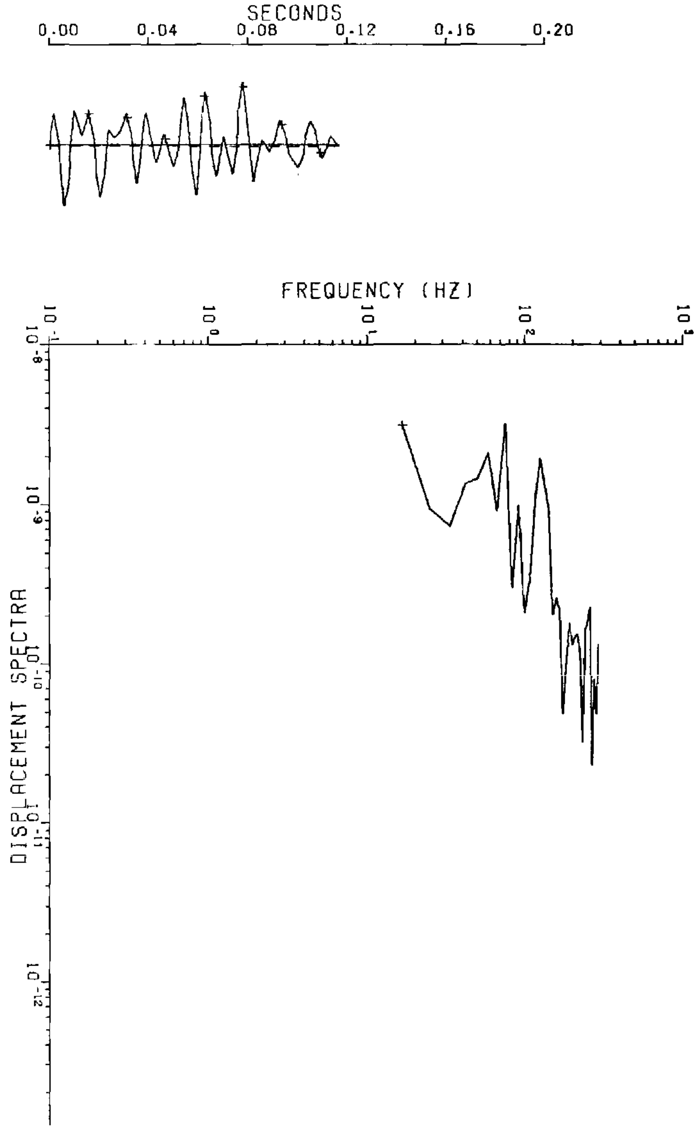
CHRA 3/26/77 16:45:40 P-WAVE #13



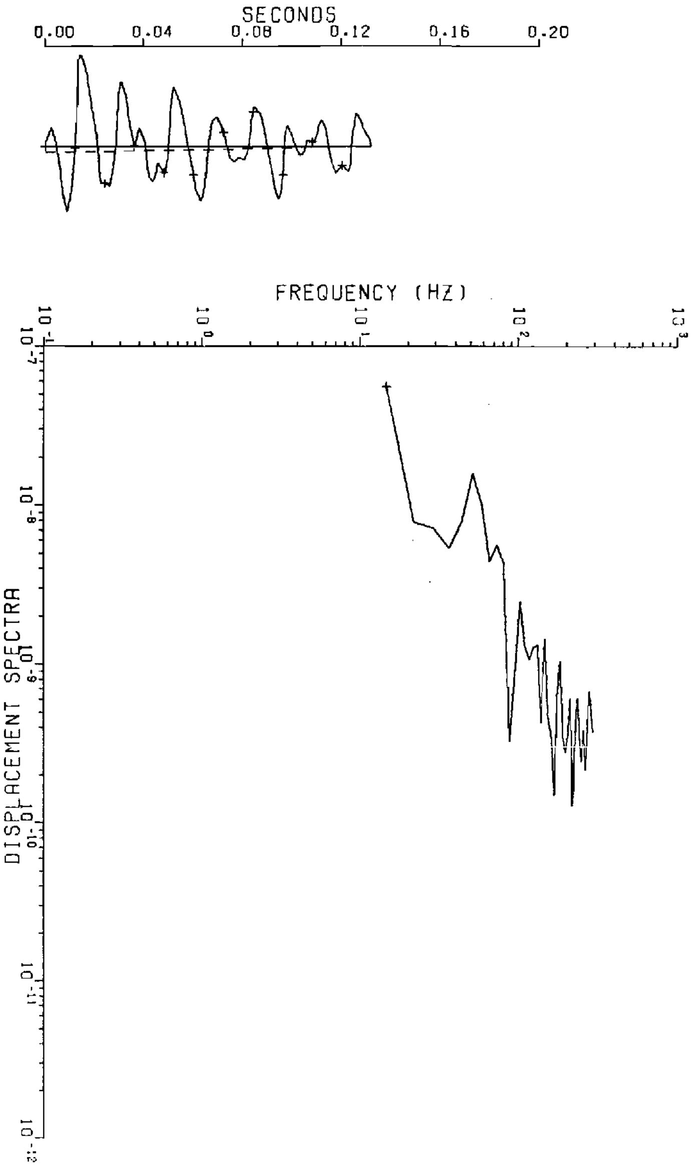
CHRA 3/26/77 16:45:40 #13



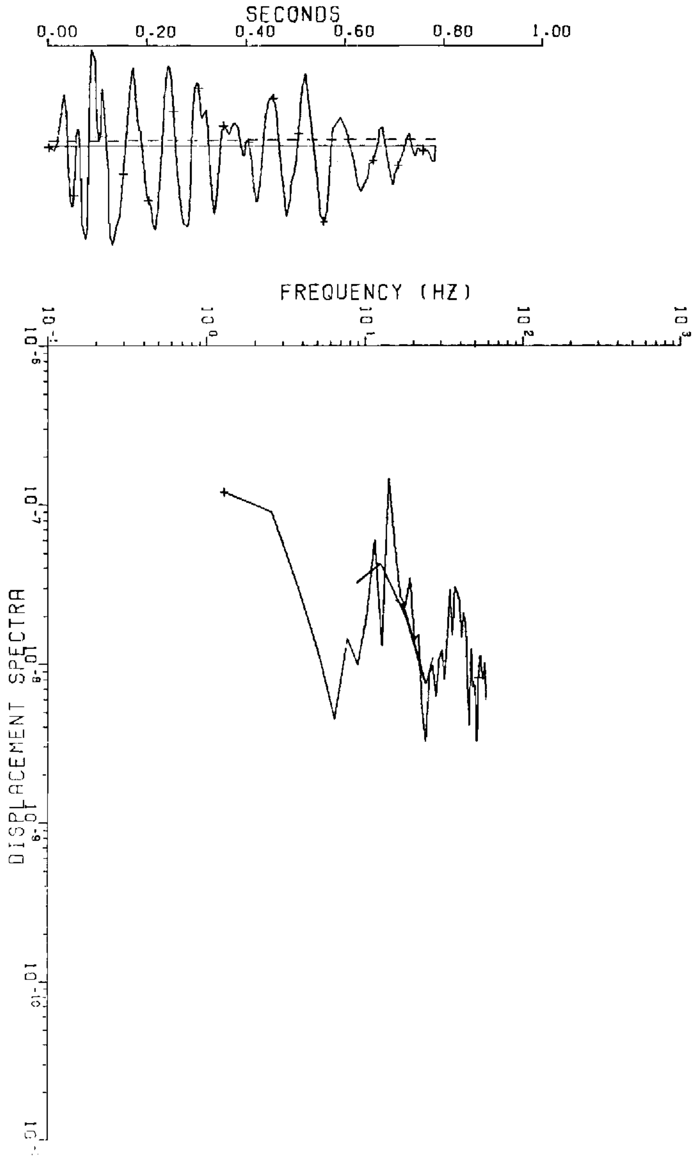
CHRA 4/14/77 16:47:17 P #14



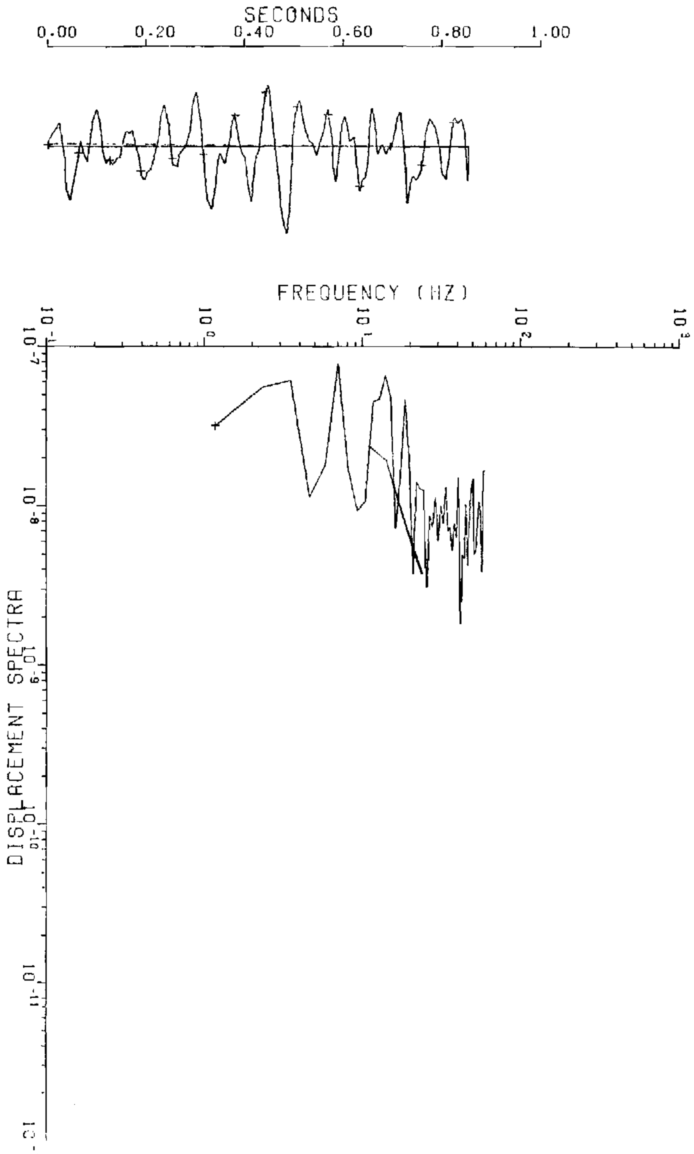
CHRA 4/14/77 16:47:17 S-WAVE #14



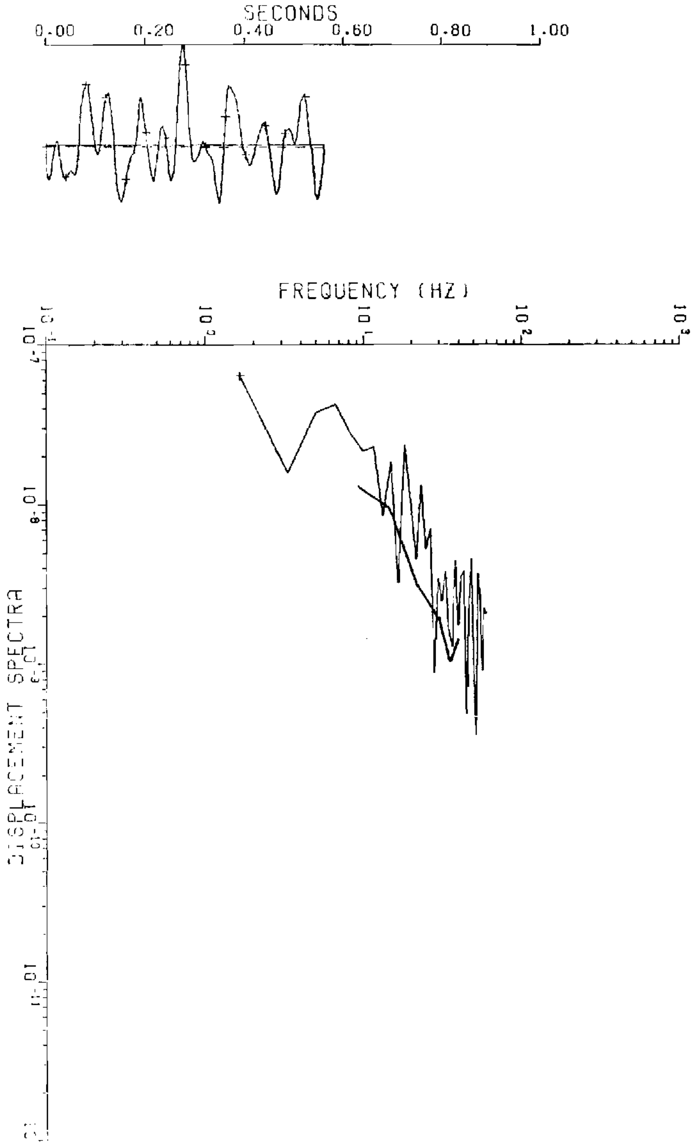
CHRA 9/24/76 19:05 CH6 P #15



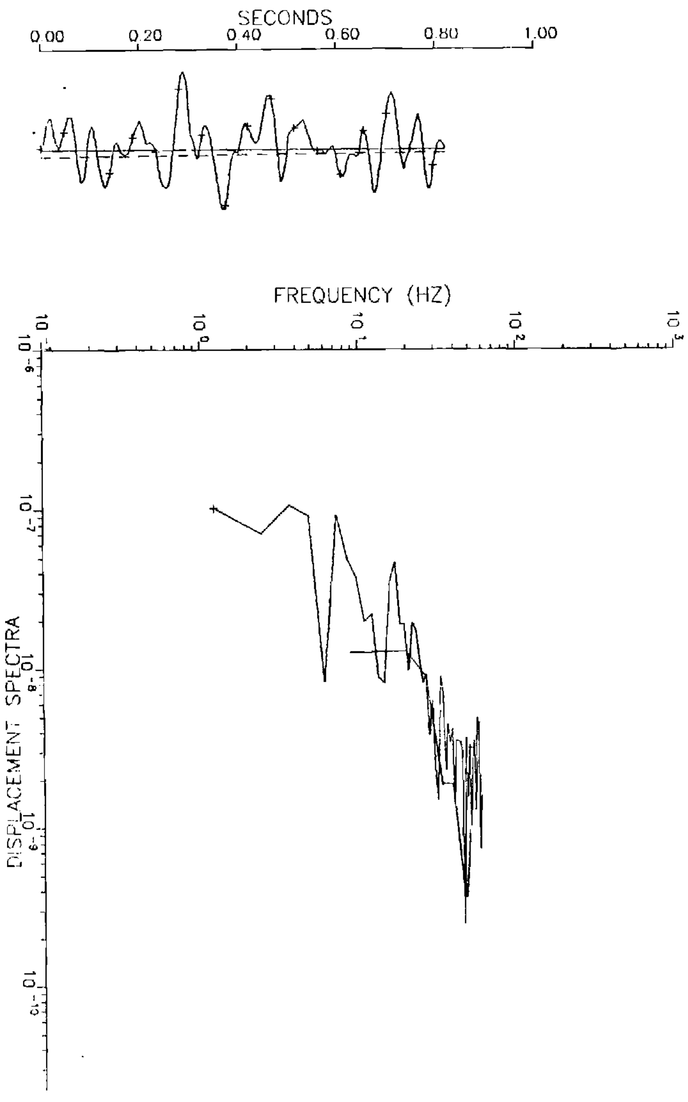
CHRA 9/29/77 09:48 M1.4 P #16



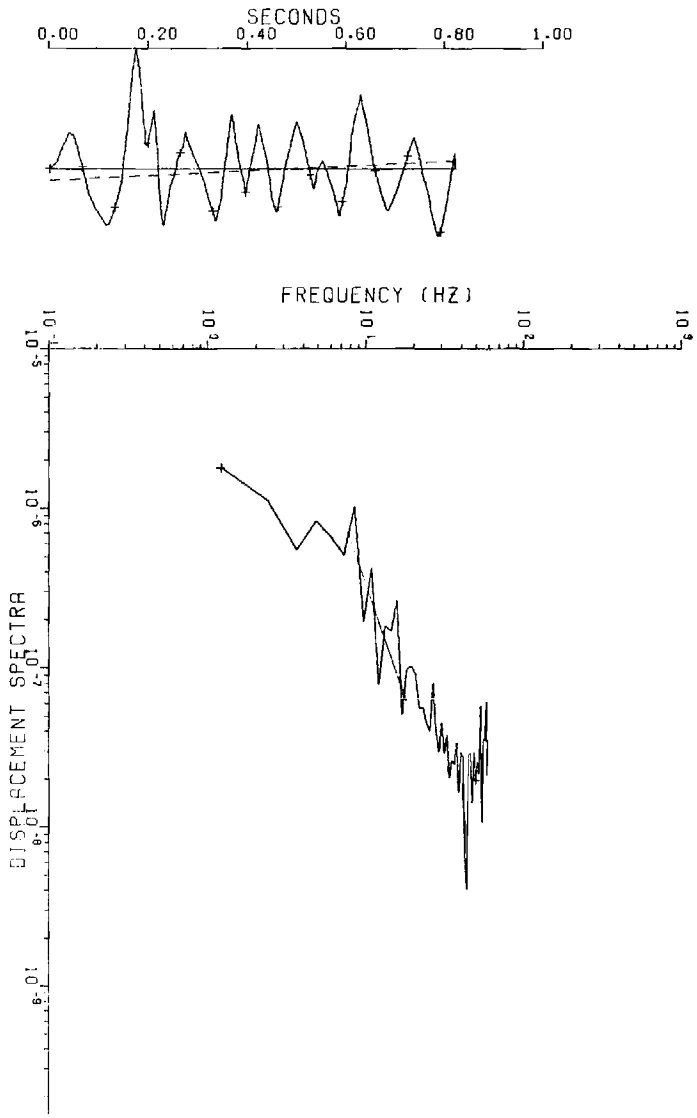
CHRA 6/12/78 06:33 CH5 #17



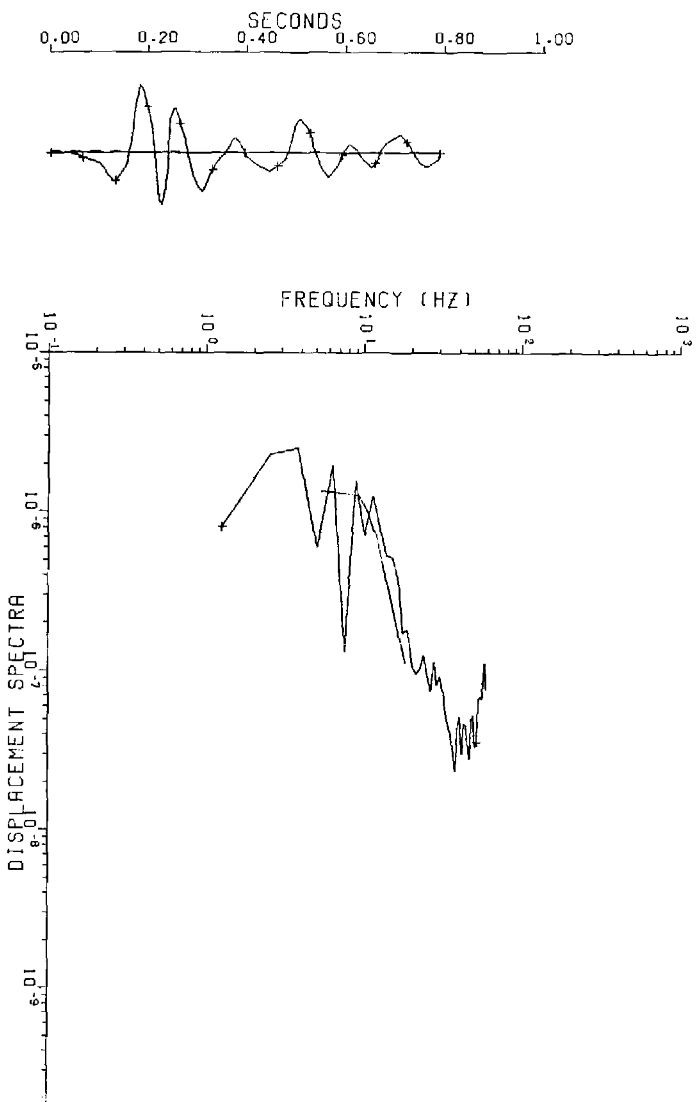
CHRA 6/12/78 06:33 CH5 S #17



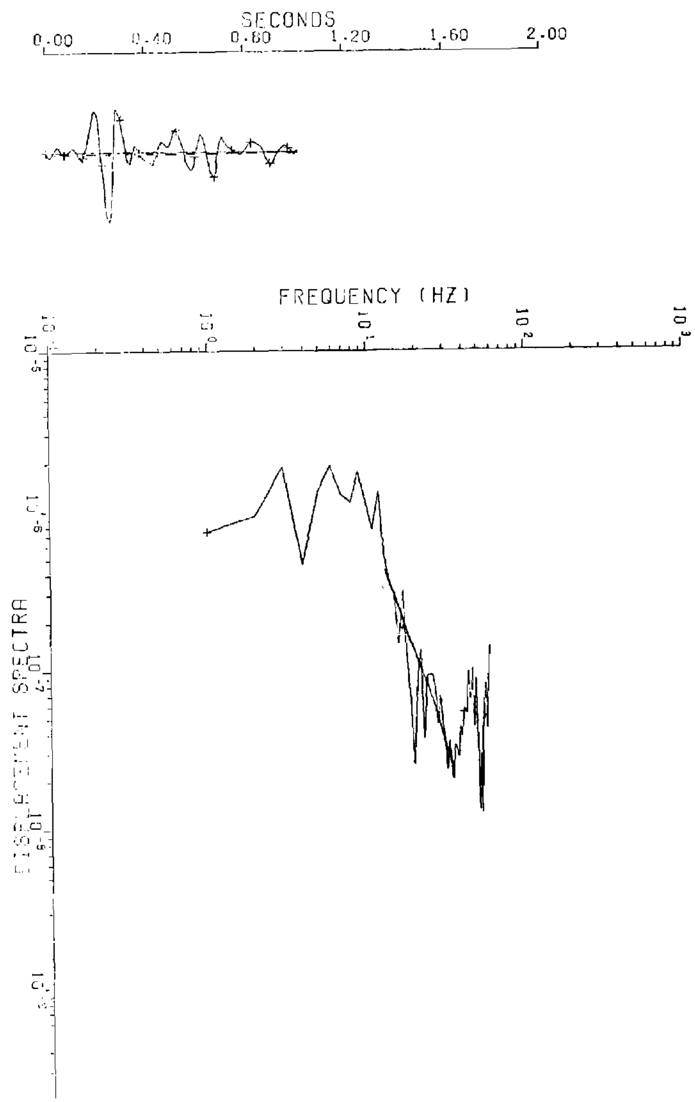
MRA 8/27/78 M=2.7 ACC Z #18



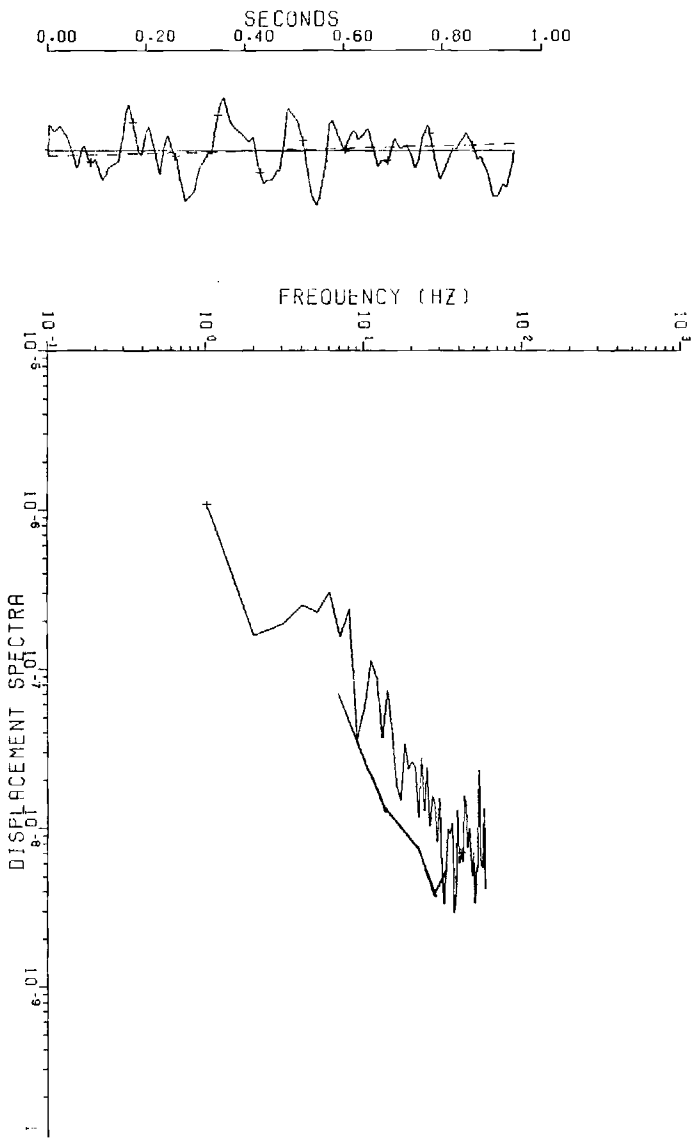
MRA 8/27/78 M=2.7 ACC NS #18



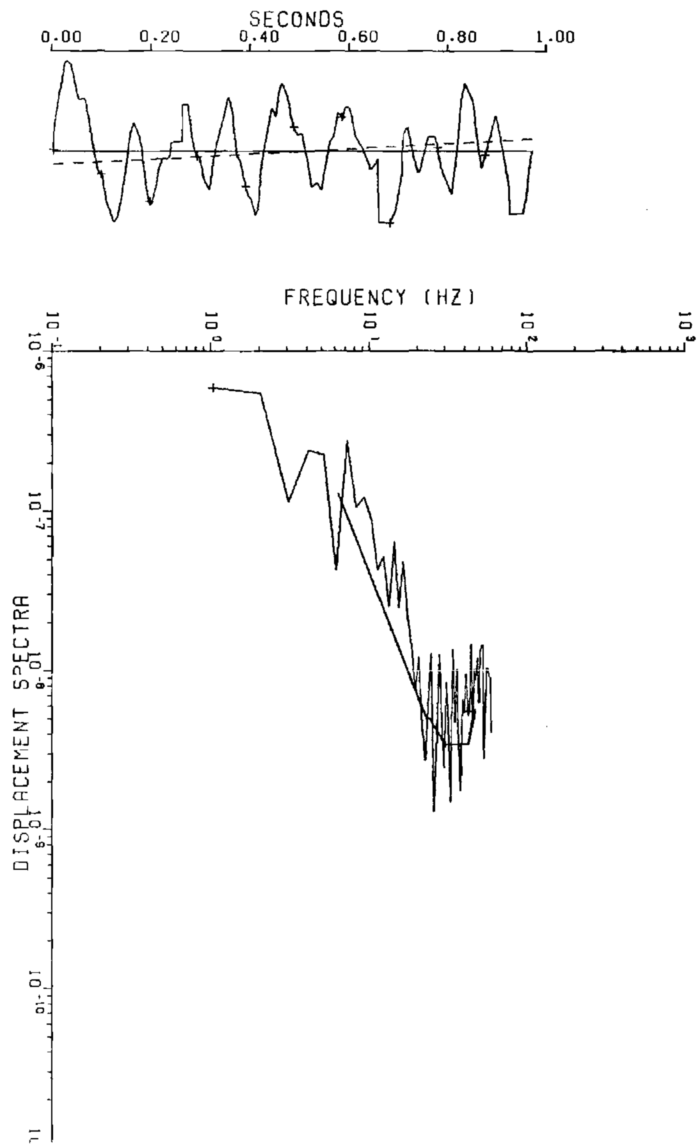
MRA 8/27/78 M=2.7 ACCEL EW #18



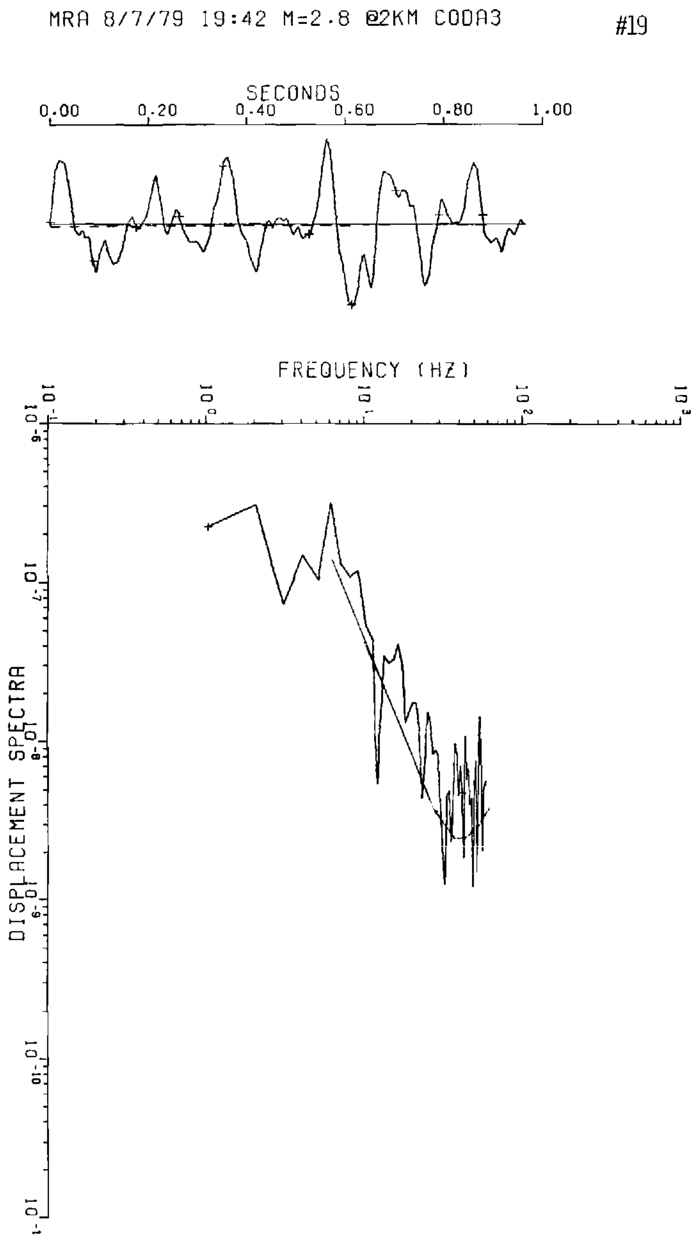
MRA 8/7/79 19:42 CODA1 #19



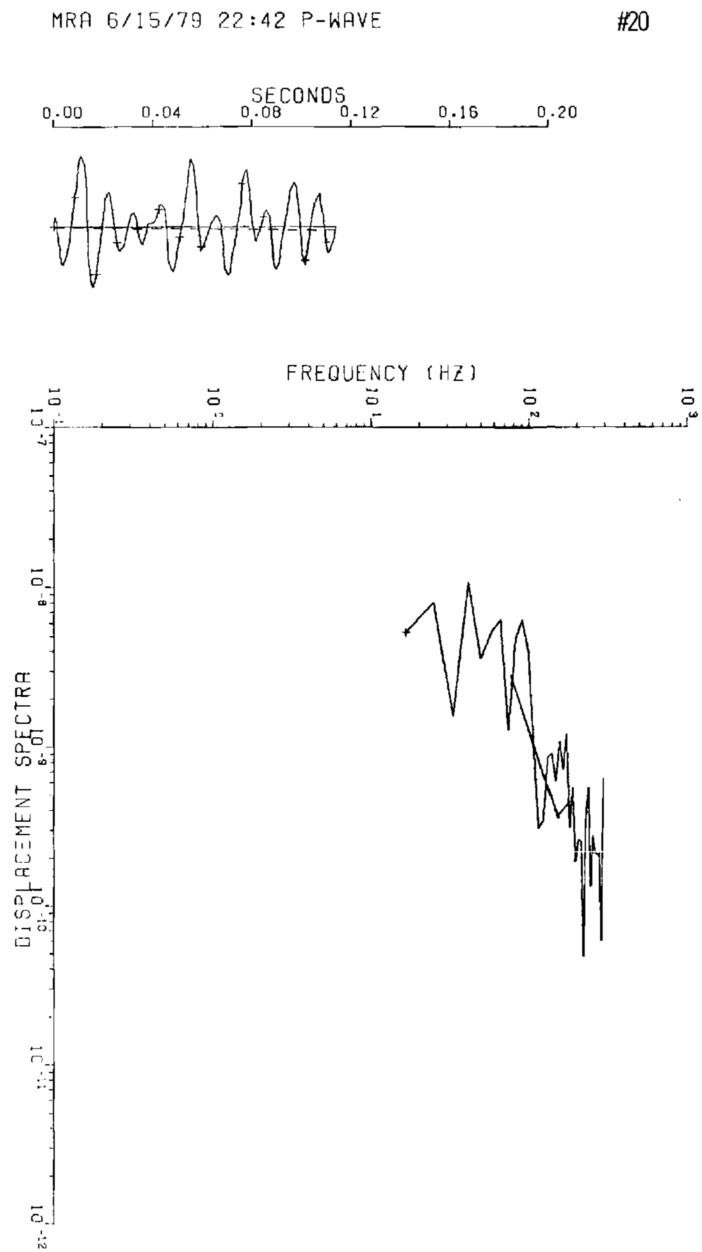
MRA 8/7/79 19:42 M=2.8 @2KM CODA2 #19



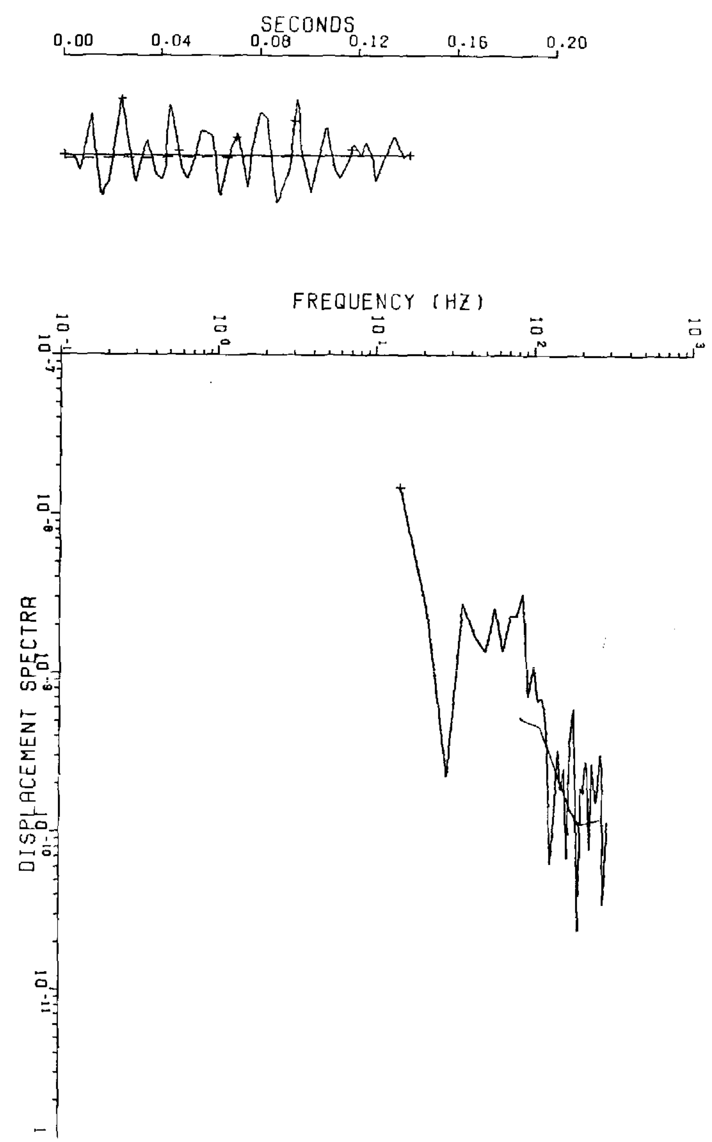
#19


MRA 8/7/79 19:42 M=2.8 @2KM CODA3
SECONDS
0.00
0.20
0.40
0.60
0.80
1.00
FREQUENCY (HZ)
DISPLACEMENT SPECTRA

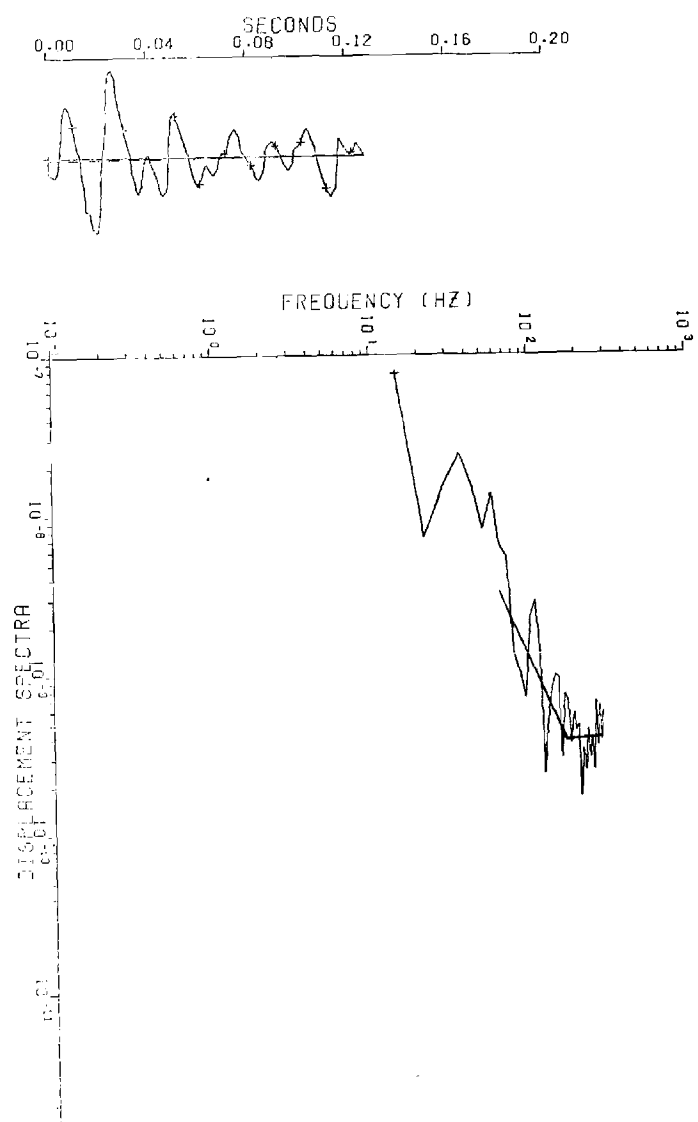
#20


MRA 6/15/79 22:42 P-WAVE
SECONDS
0.00
0.04
0.08
0.12
0.16
0.20
FREQUENCY (HZ)
DISPLACEMENT SPECTRA

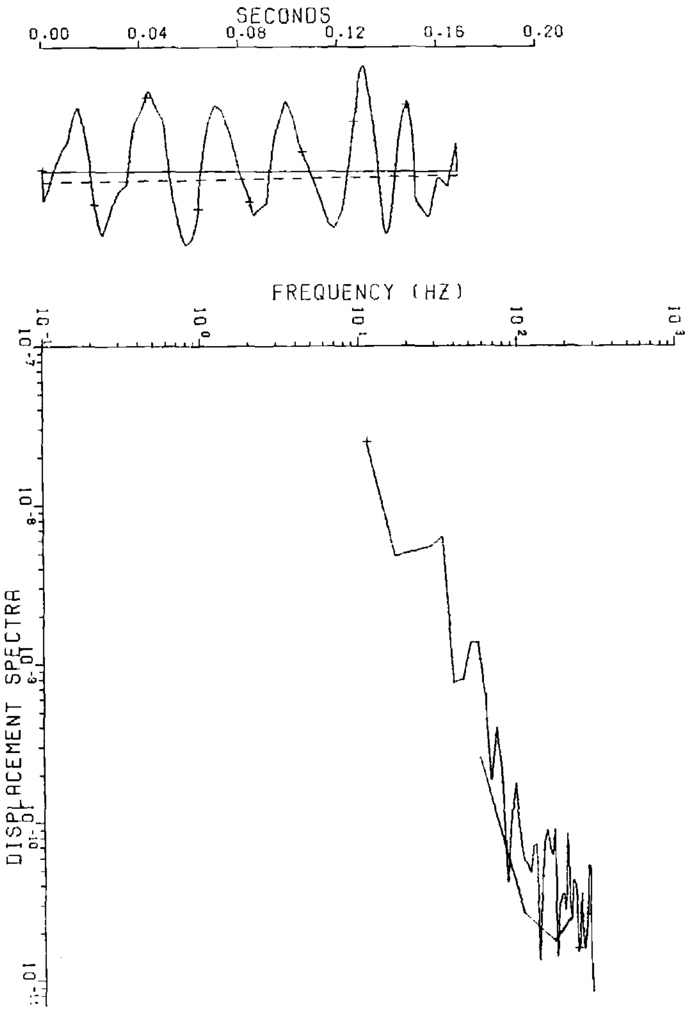
MRA 6/15/79 22:42 P CODA #20



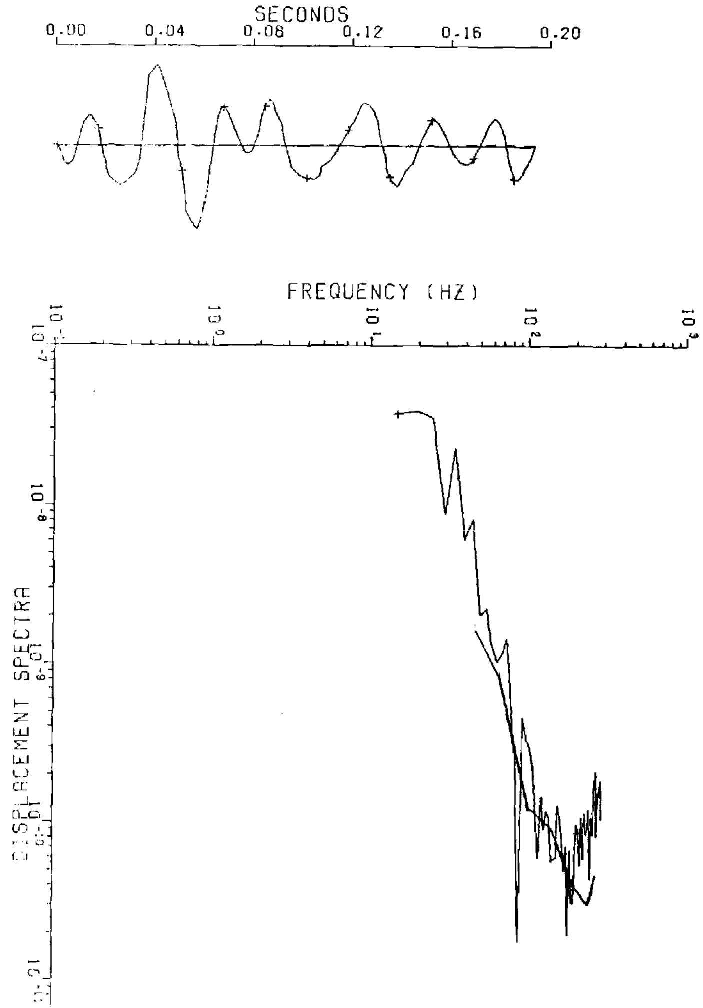
MRA 6/15/79 S-WAVE #20



MRA 11/16/79 14:29:22 P #21

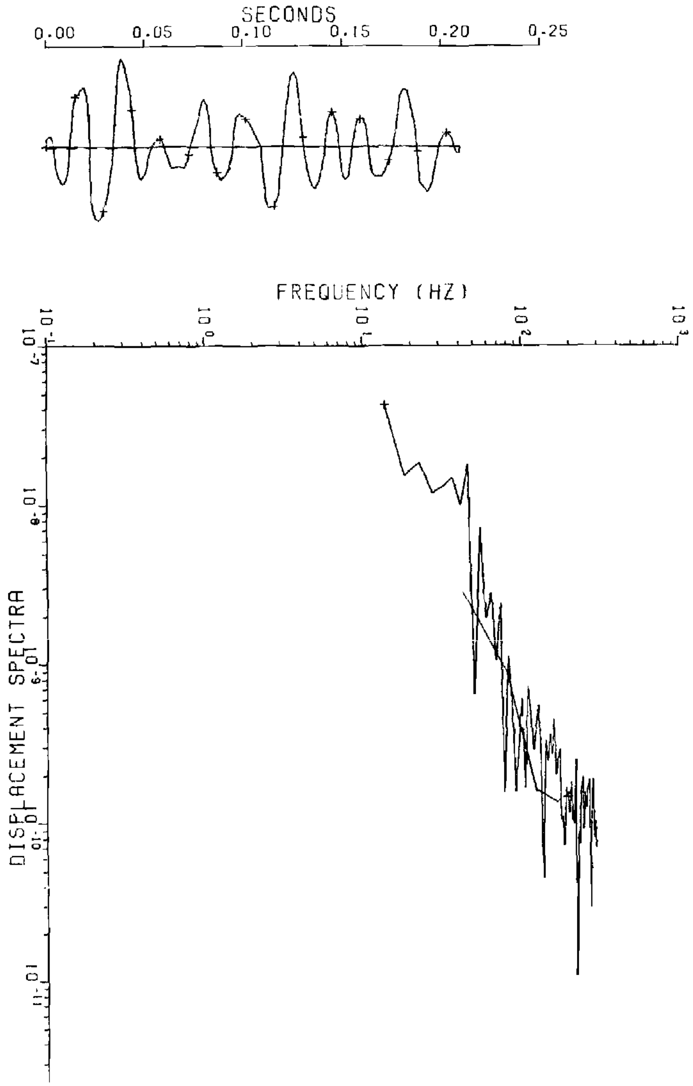


MRA 11/16/79 14:29 S #21



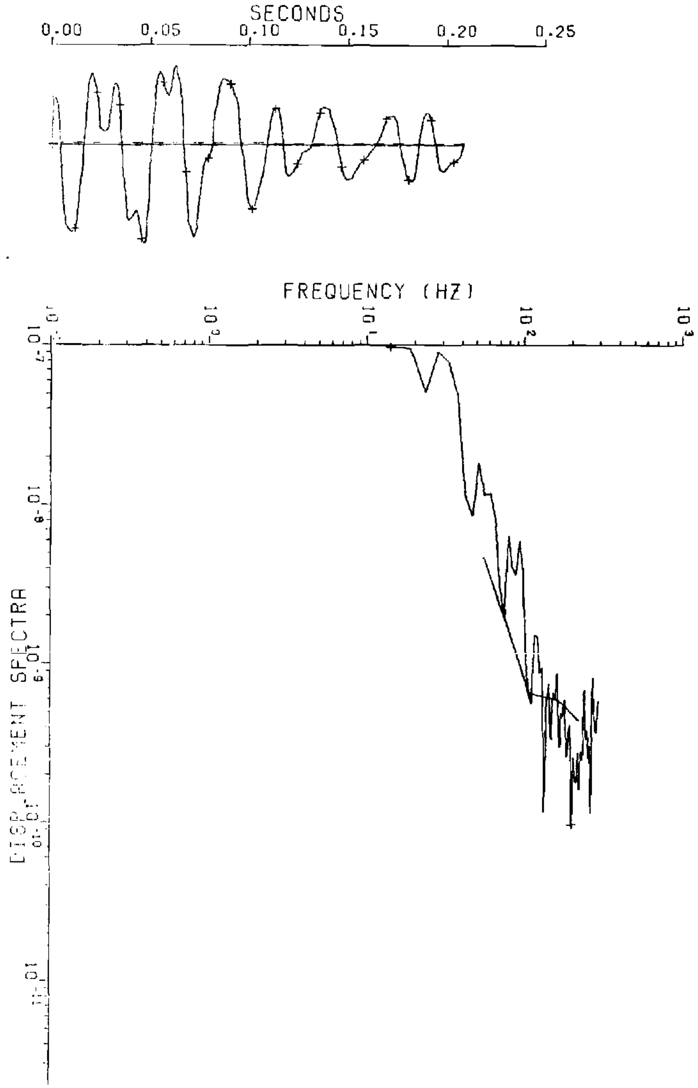
MRA 11/16/79 15:15 P

#22

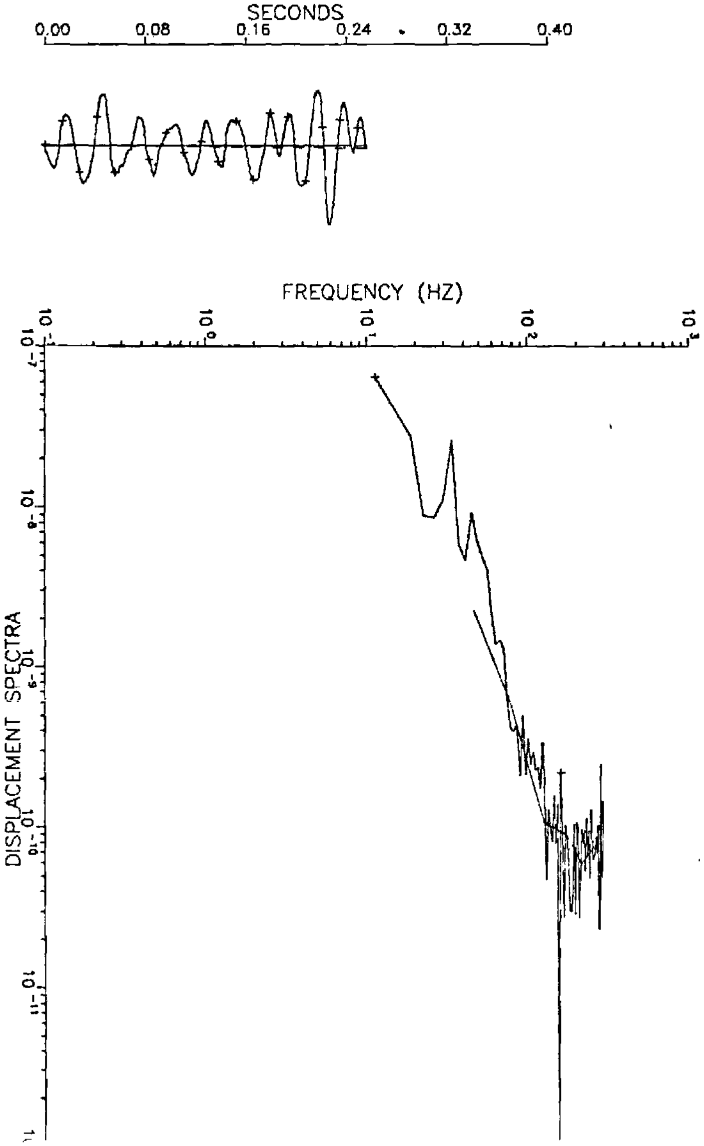


MRA 11/16/79 15:15 S

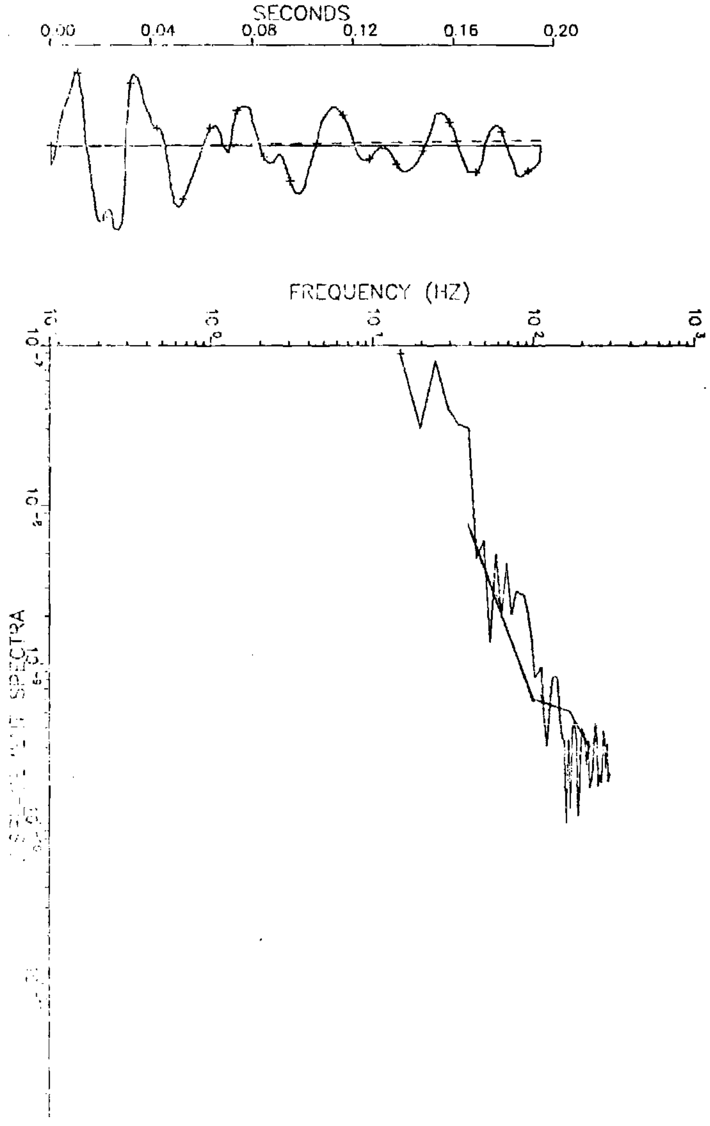
#22

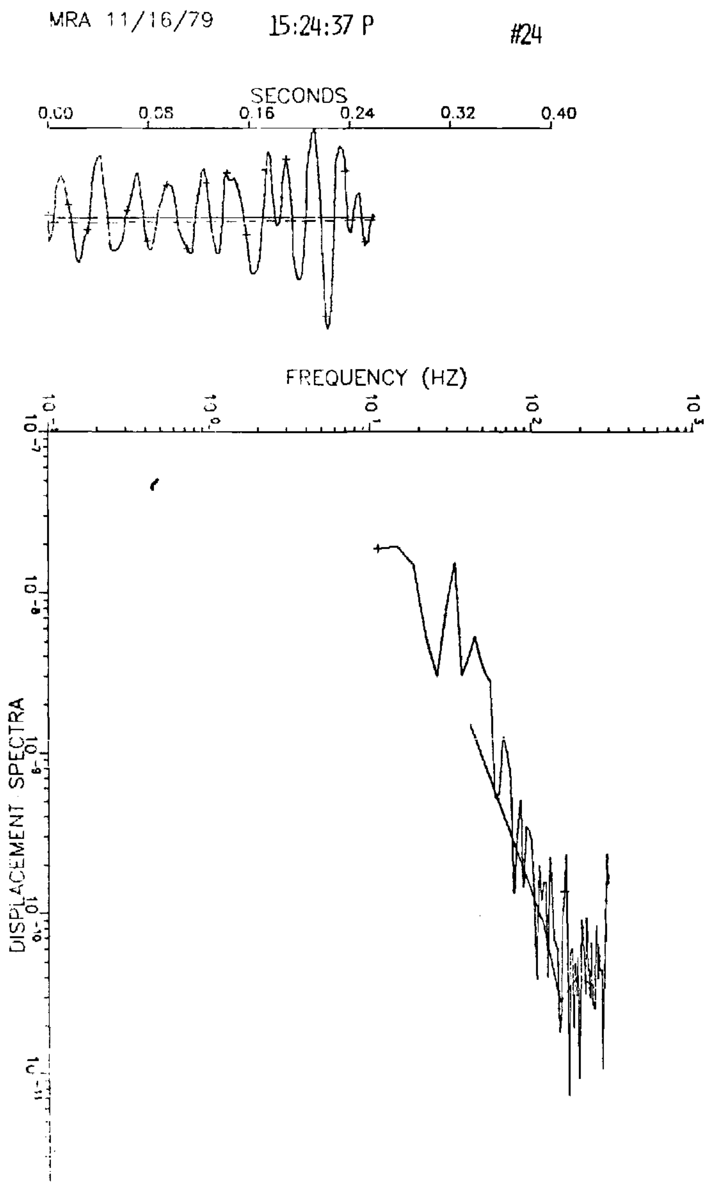


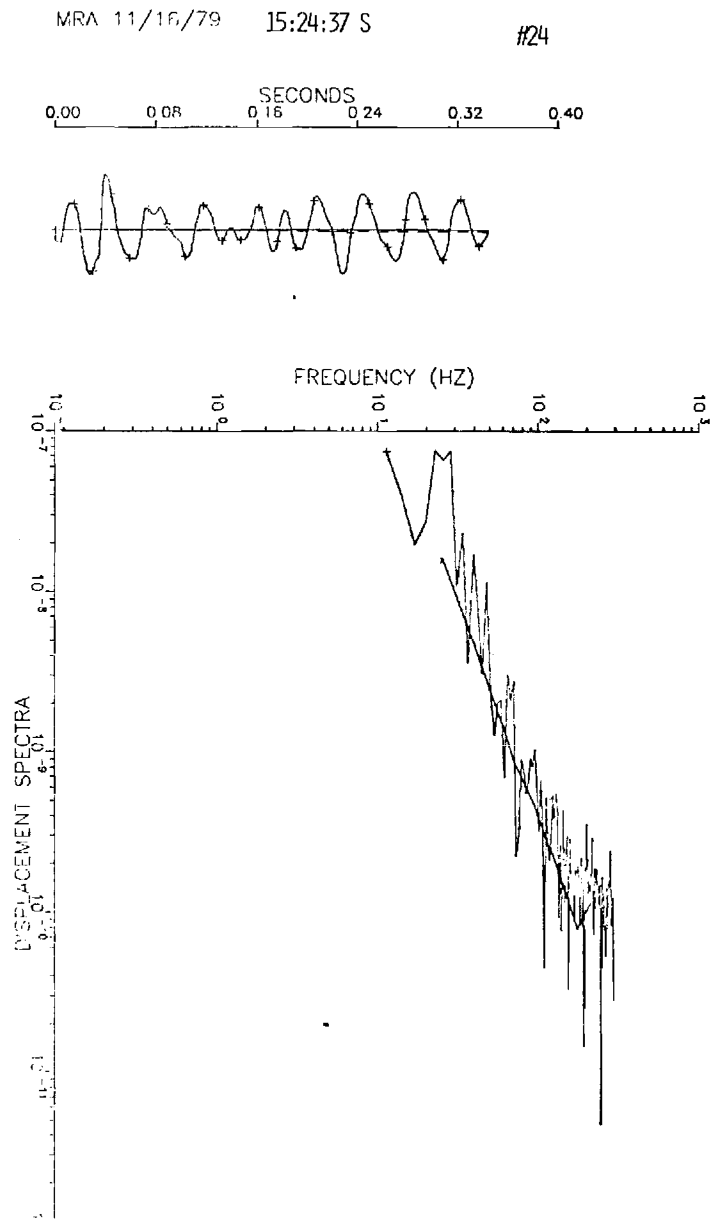
MRA 11/16/79 15:24:21 P #23



MRA 11/16/79 15:24:21 S #23

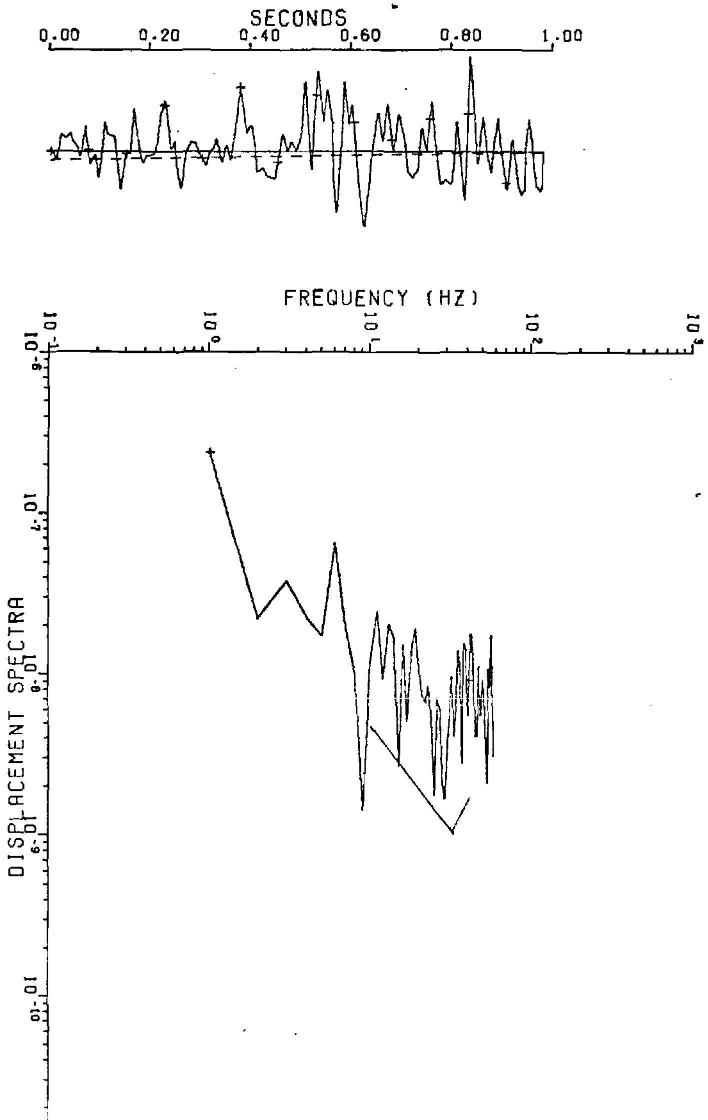



MRA 11/16/79
15:24:37 P
#24
SECONDS
0.00
0.08
0.16
0.24
0.32
0.40
FREQUENCY (HZ)
DISPLACEMENT SPECTRA


MRA 11/16/79
15:24:37 S
#24
SECONDS
0.00
0.08
0.16
0.24
0.32
0.40
FREQUENCY (HZ)
DISPLACEMENT SPECTRA

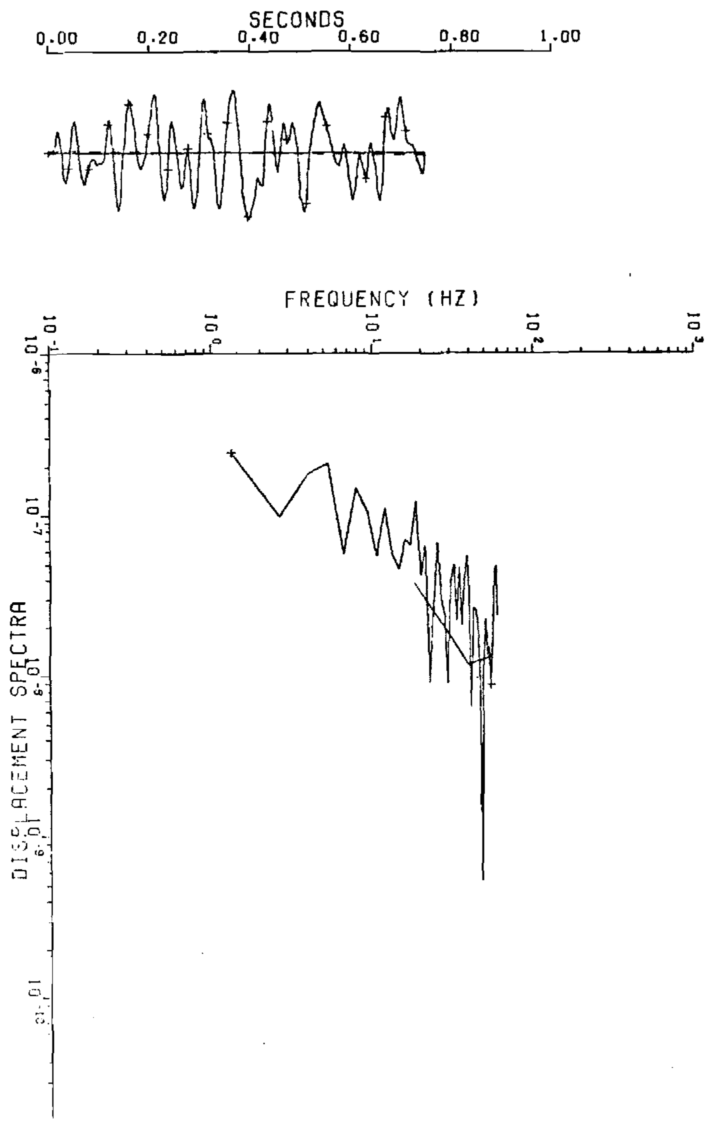
LSA 8/31/77 16:56 ETG P

#25



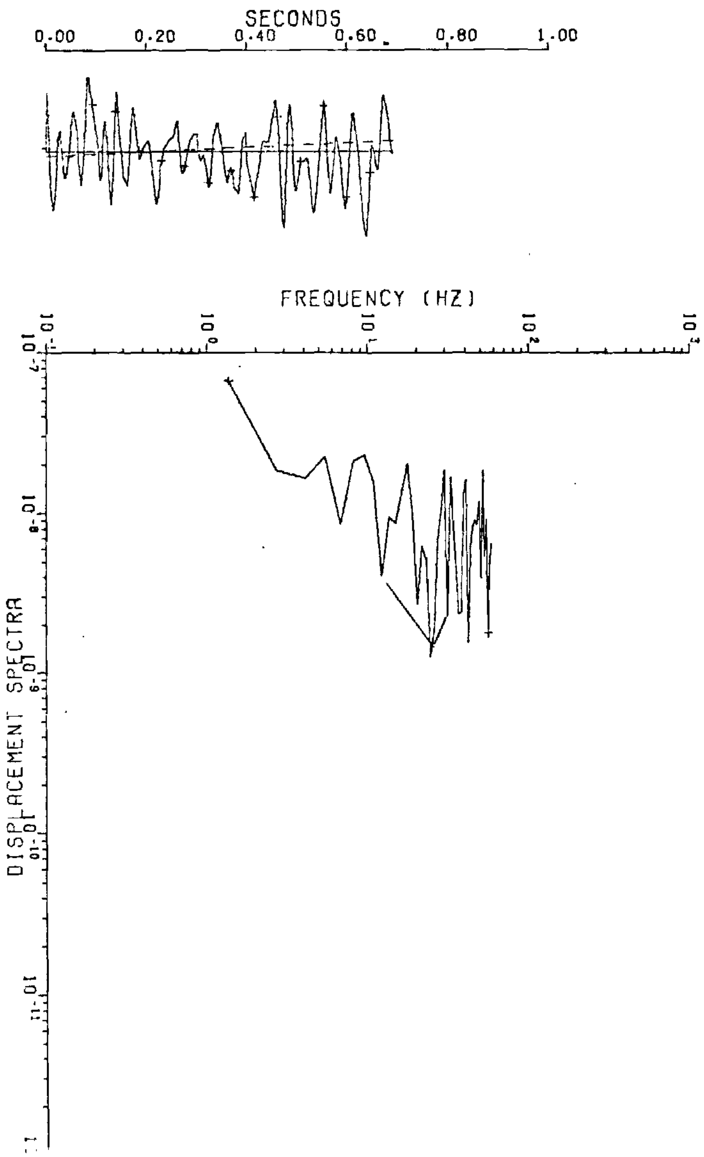
LSA 8/31/77 ETG S

#25



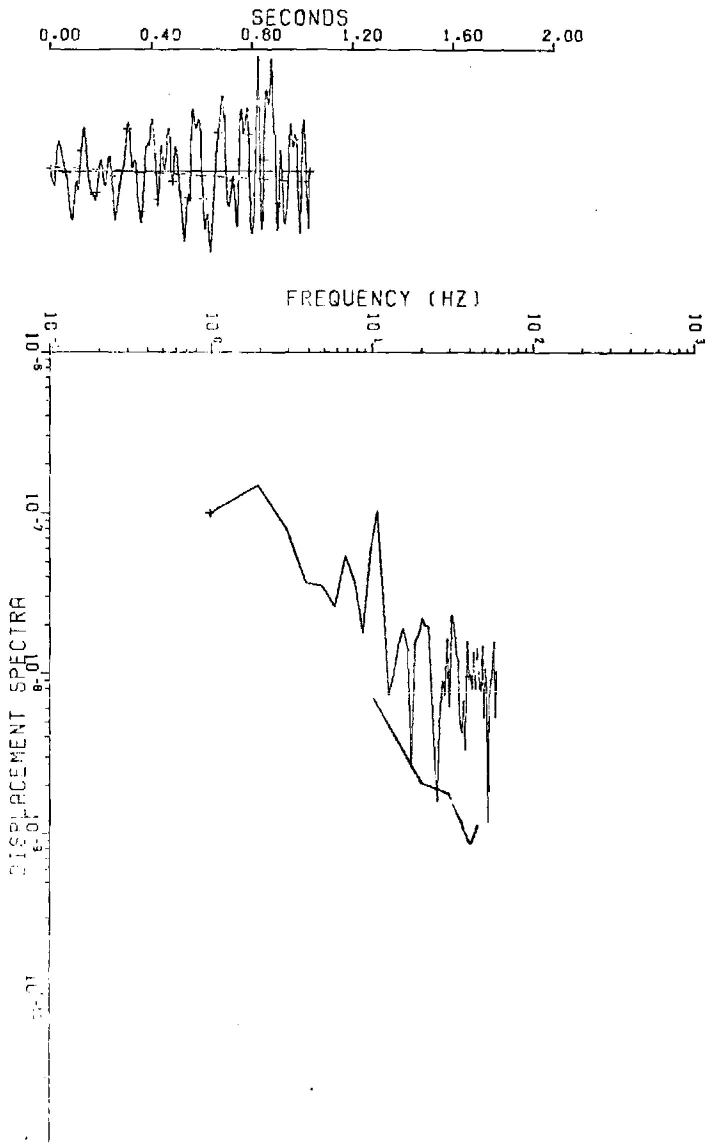
LSA 8/31/77 16:56 REG P

#25



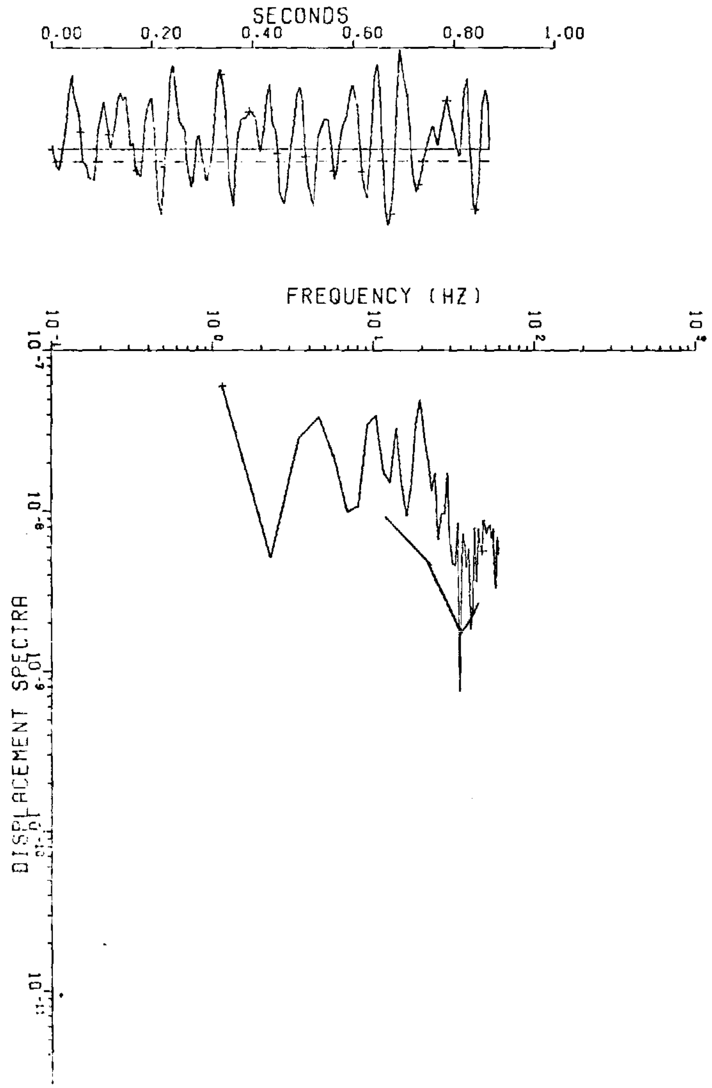
LSA 9/6/77 20:04 REG P

#26



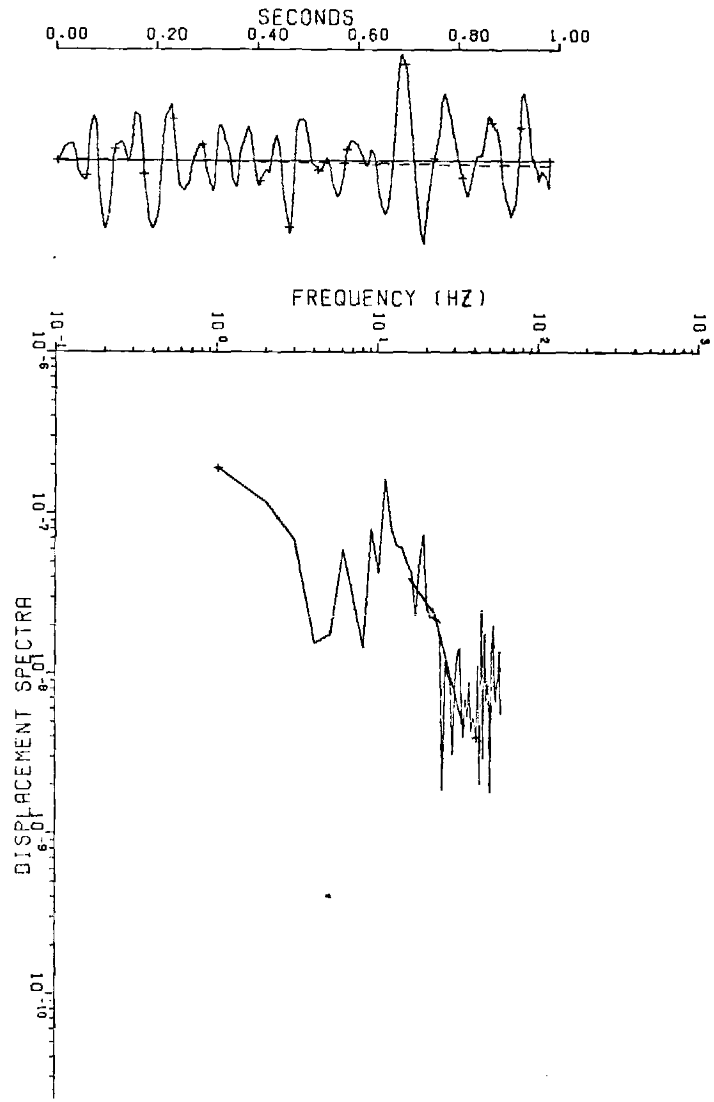
LSA 9/6/77 20:04 GBG P

#26



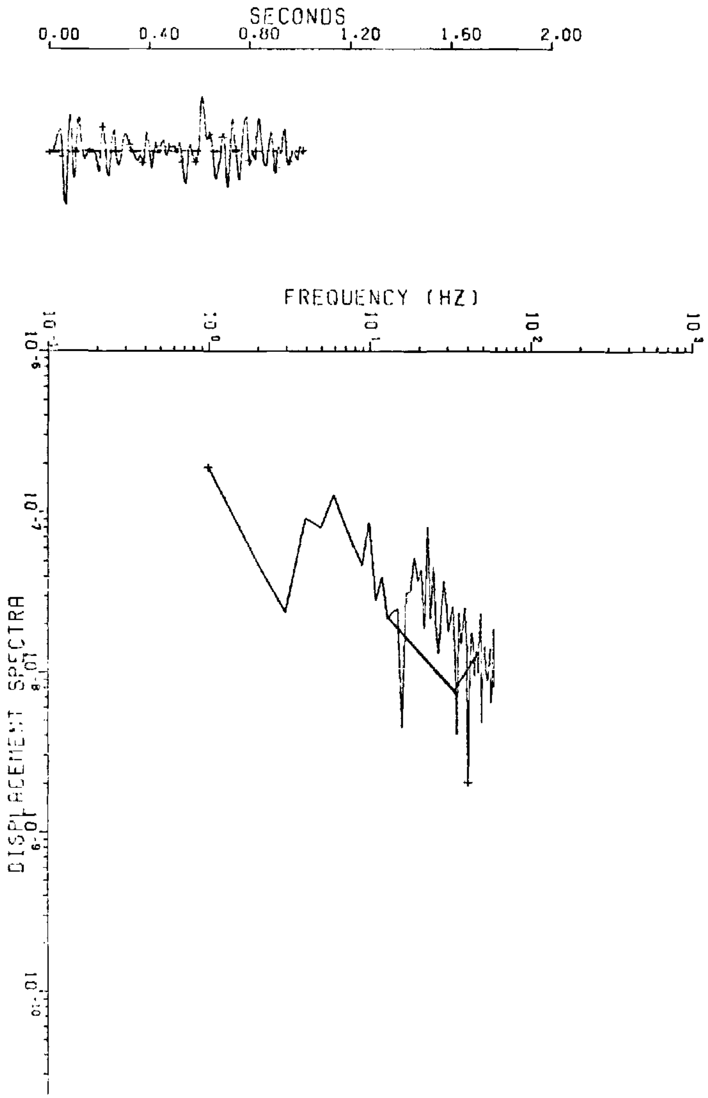
LSA 9/6/77 20:04 GBG S

#26



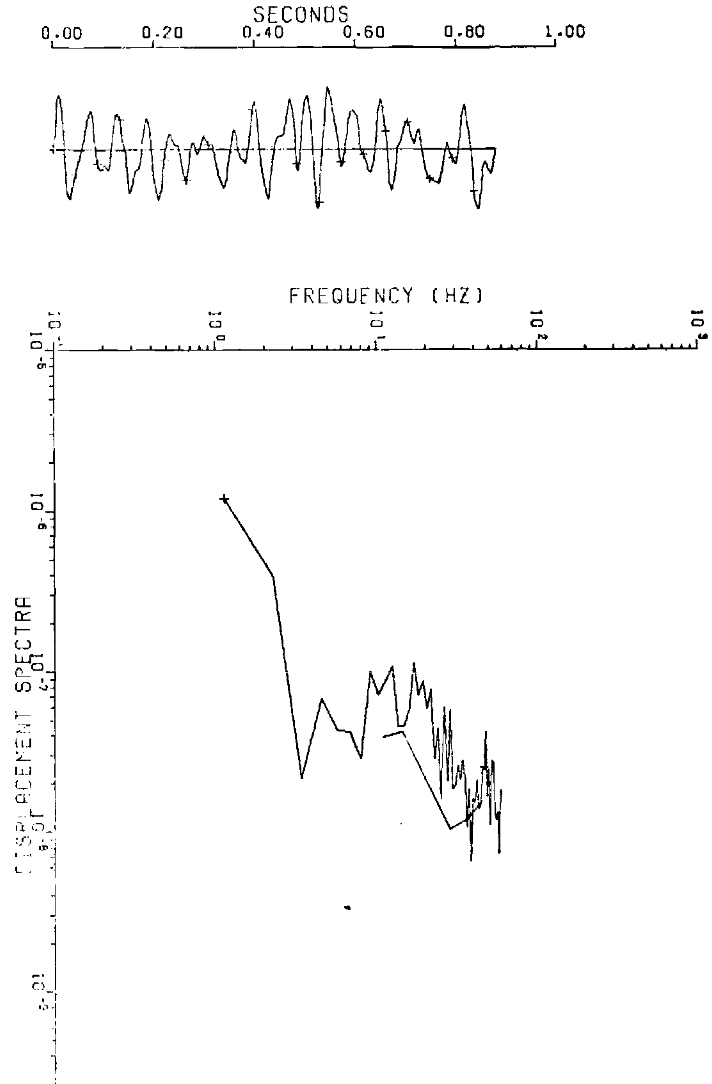
LSA 11/23/77 22:30 GBG P

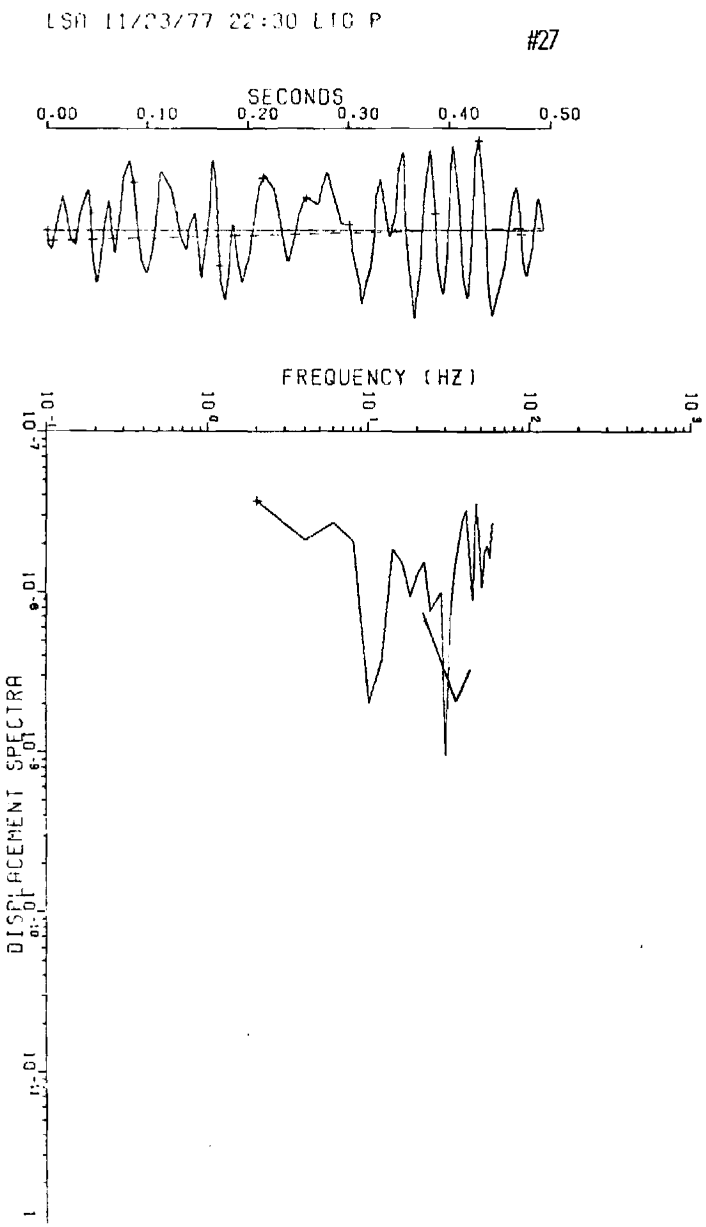
#27

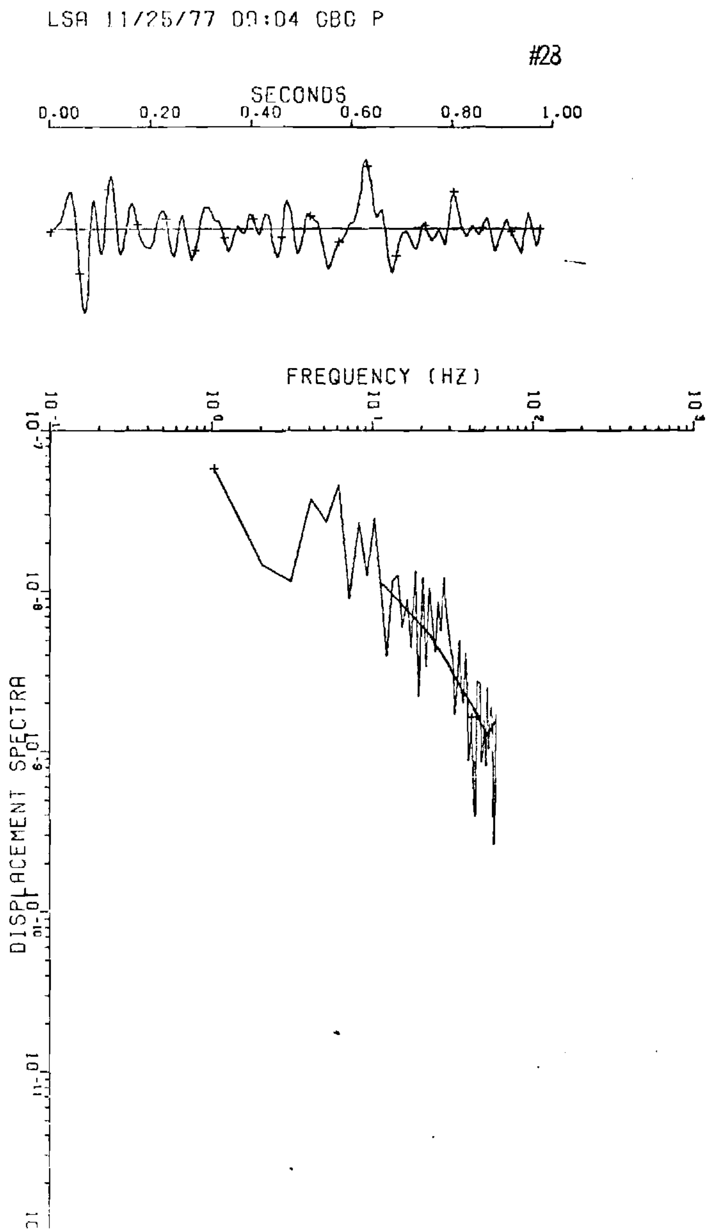


LSA 11/23/77 22:30 GBG S

#27

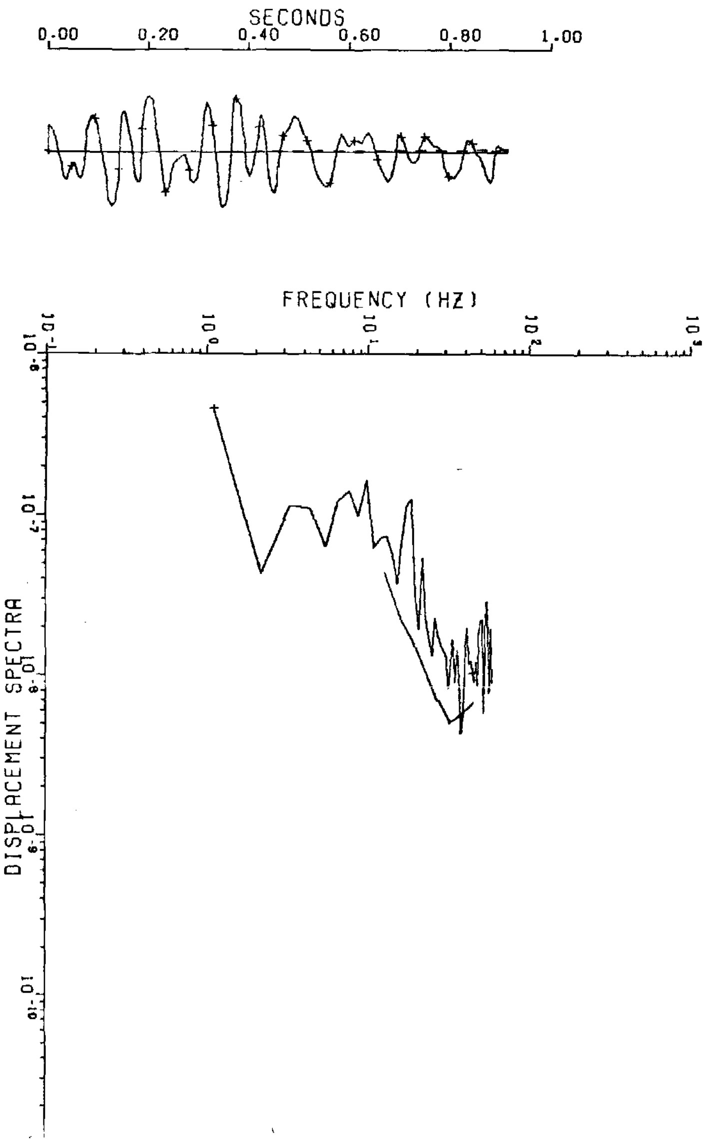



LSA 11/23/77 22:30 LTC P
#27
SECONDS
0.00 0.10 0.20 0.30 0.40 0.50
FREQUENCY (HZ)
DISPLACEMENT SPECTRA


LSA 11/25/77 09:04 GBG P
#28
SECONDS
0.00 0.20 0.40 0.60 0.80 1.00
FREQUENCY (HZ)
DISPLACEMENT SPECTRA

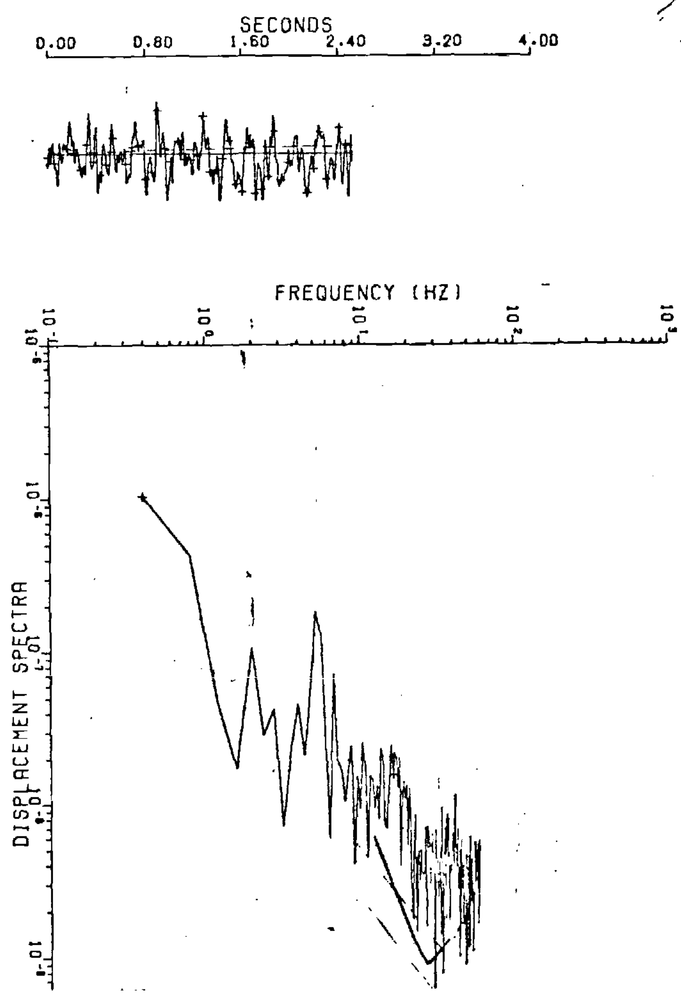
LSA 11/25/77 09:04 GBG S

#28



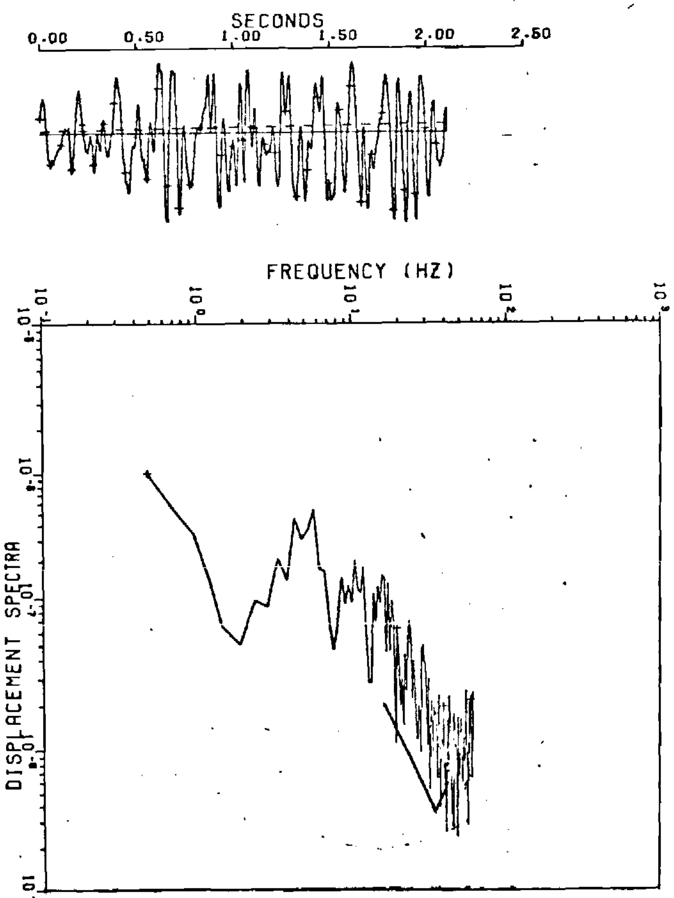
WDG-P WDA 3/20/78 34KM

#29



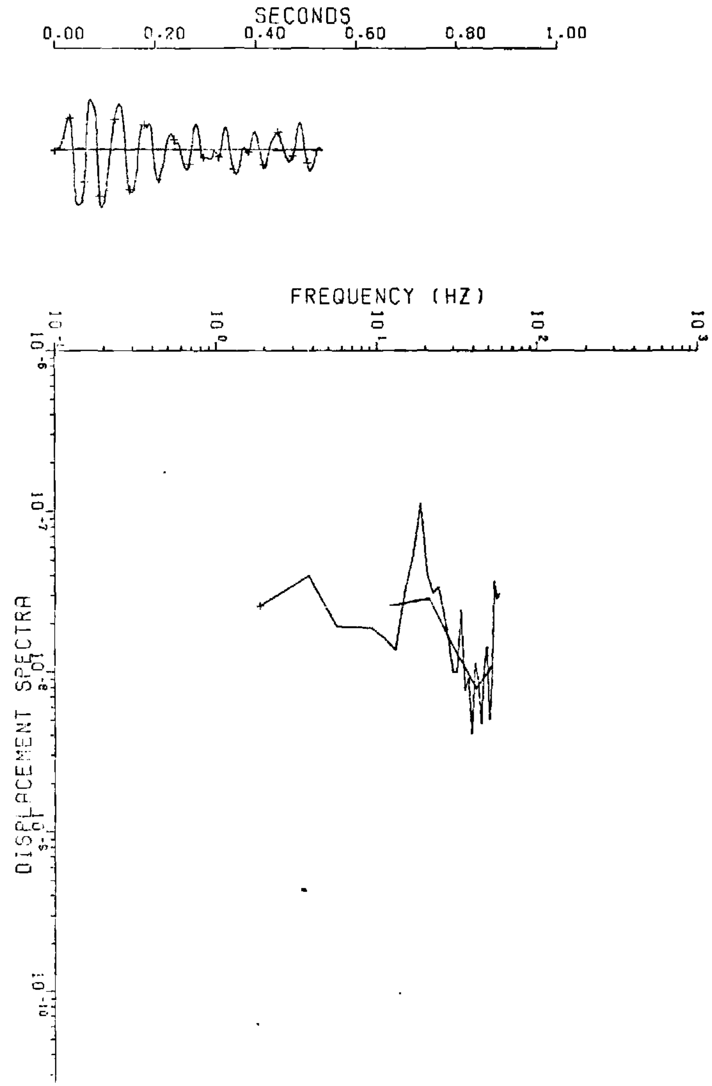
WDG-S WDA 3/20/78

#29

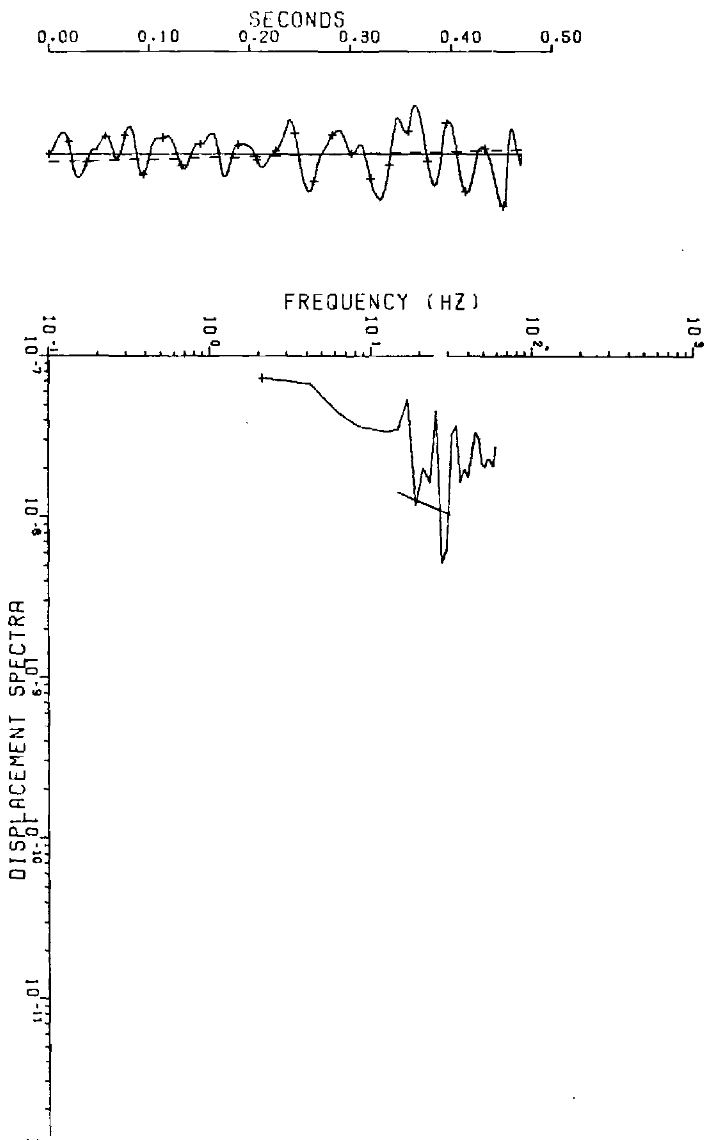


LSA 5/1/78 21:30 p

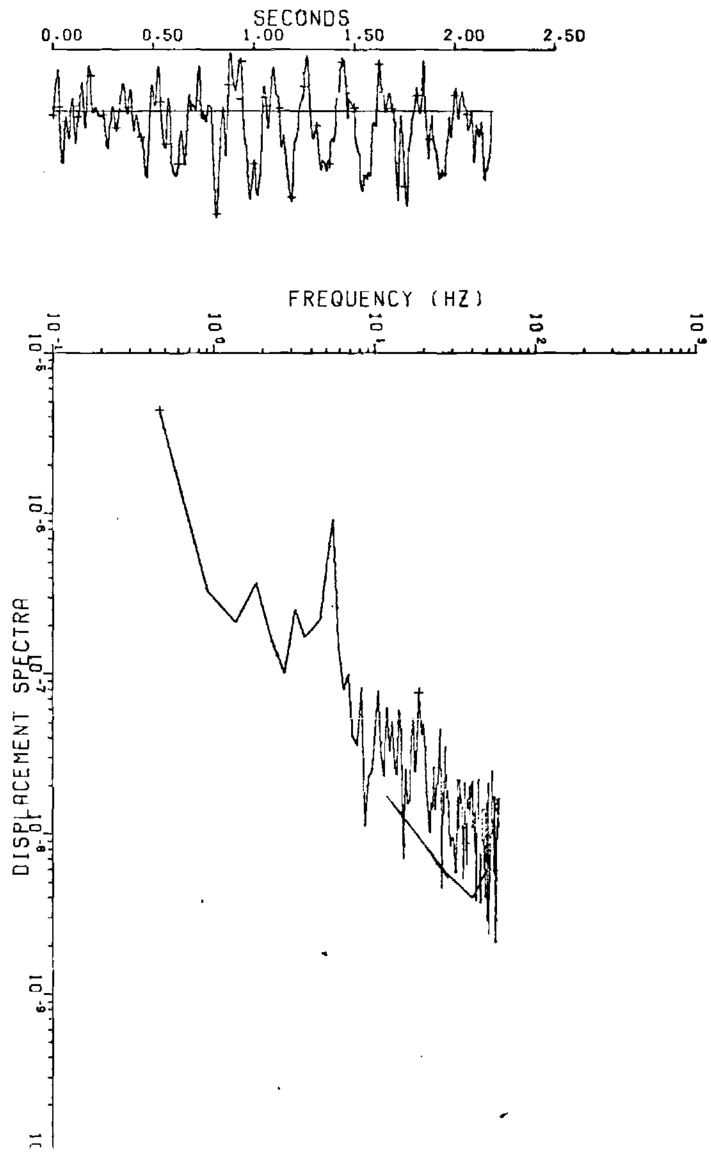
#30



LSA 5/1/78 21:30 REG S #30

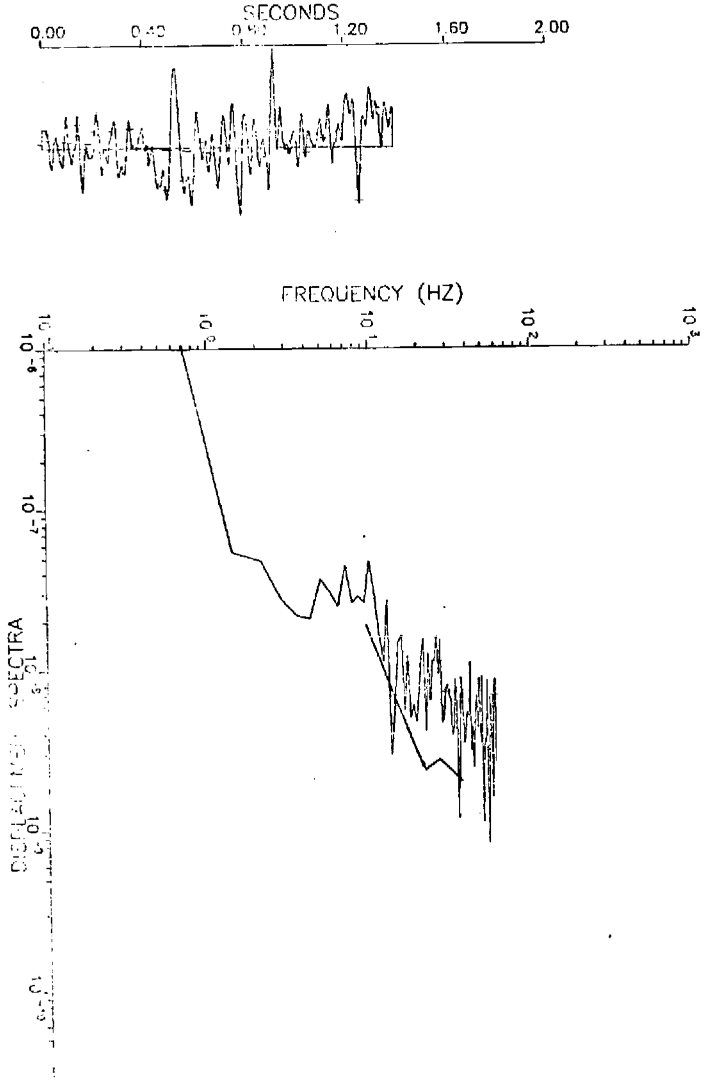


LSA 5/2/78 WDC 01:46 #31



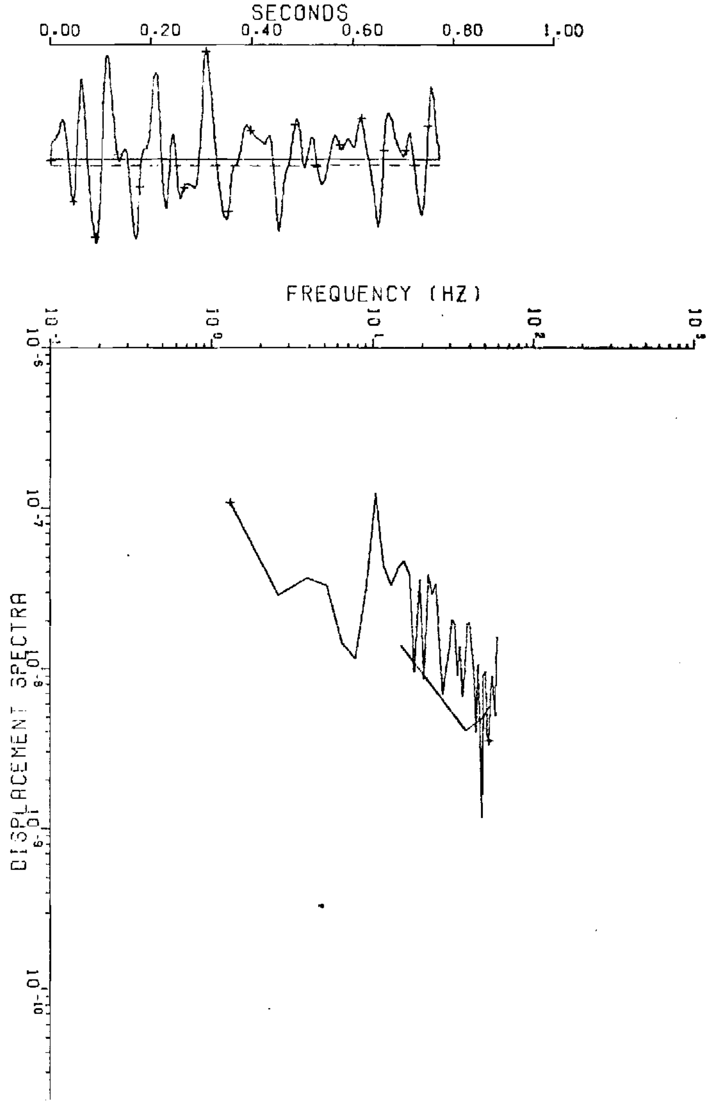
LSA 5/2/78 M=2.3 CBGP

#31



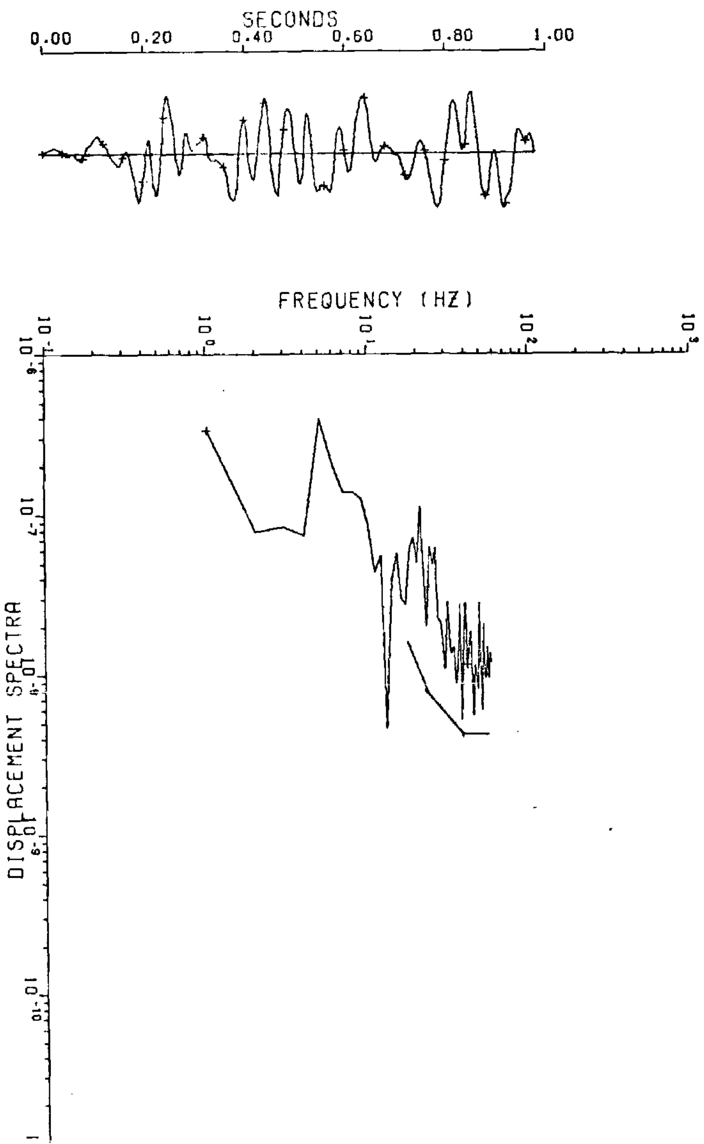
LSA 10/2/78 GBG P

#32



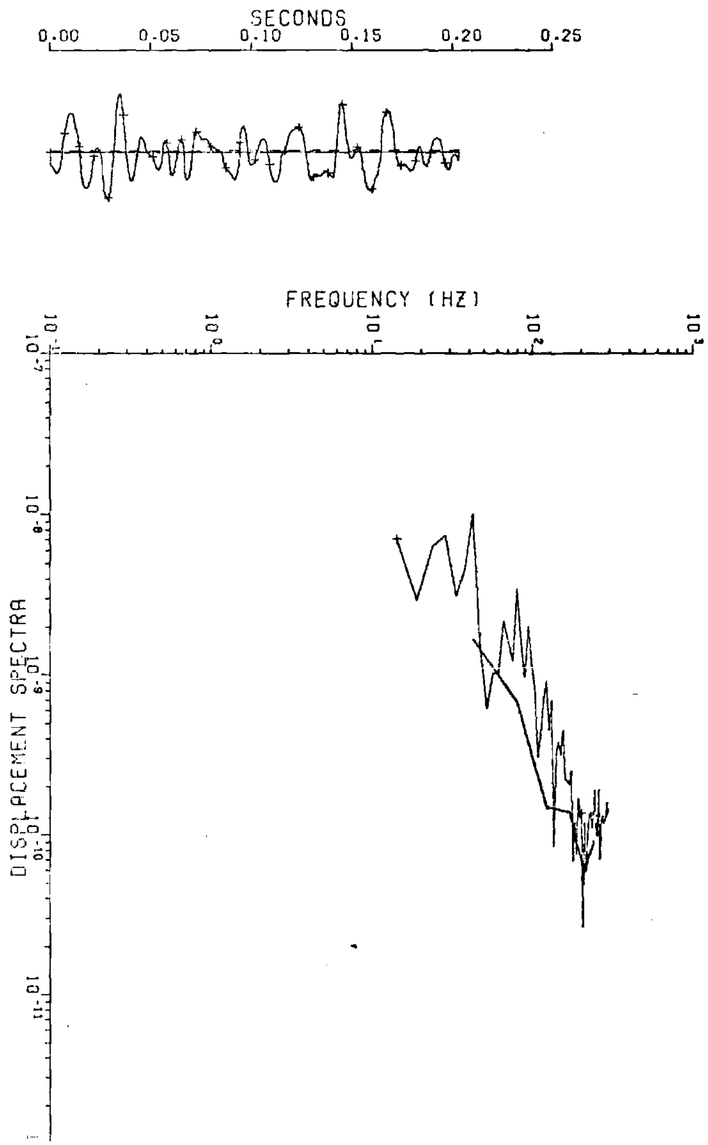
LSA 10/2/78 WDG P

#32



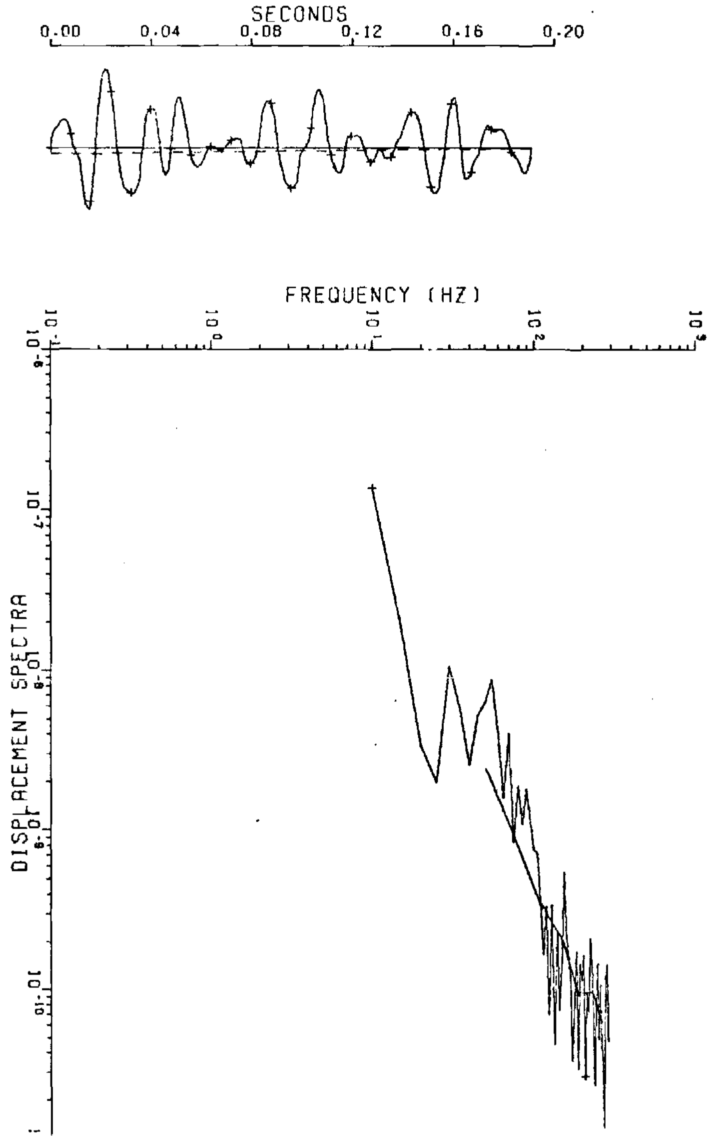
LSA 11/8/79 STATS ?23:56 P

#33

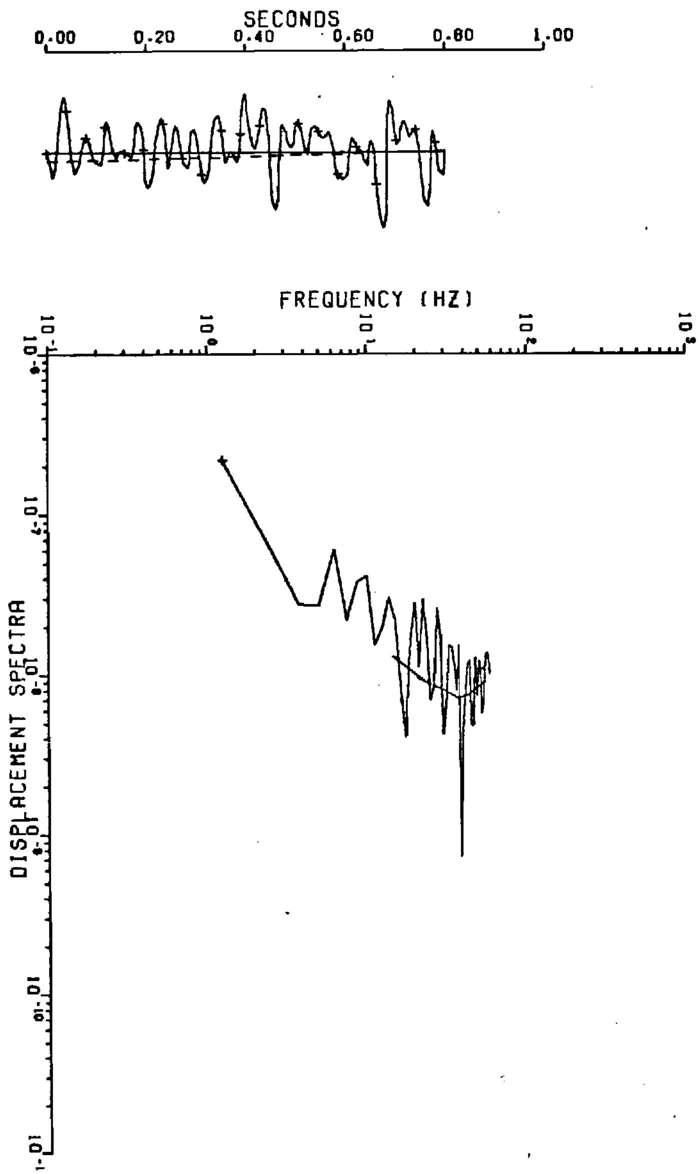


LSA 11/8/79 23:56 S

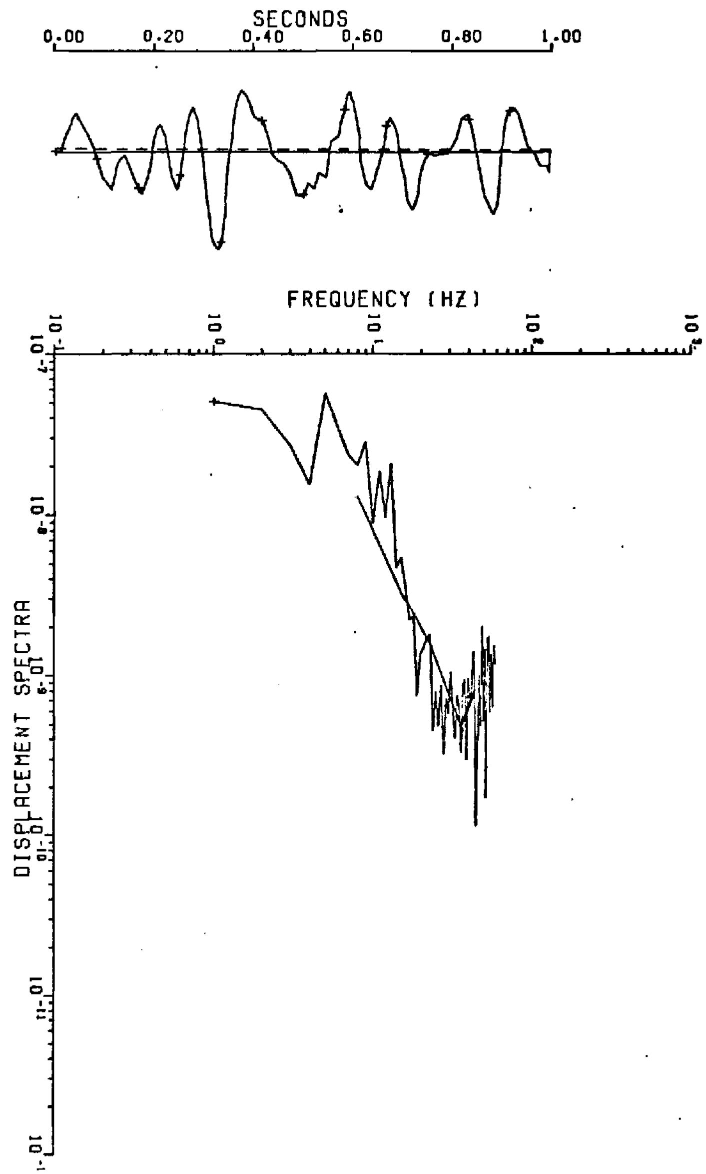
#33



MRA 8/6/79 M=2.8 CH6 PLG



MRA 8/6/79 M=2.8 GBG PLG



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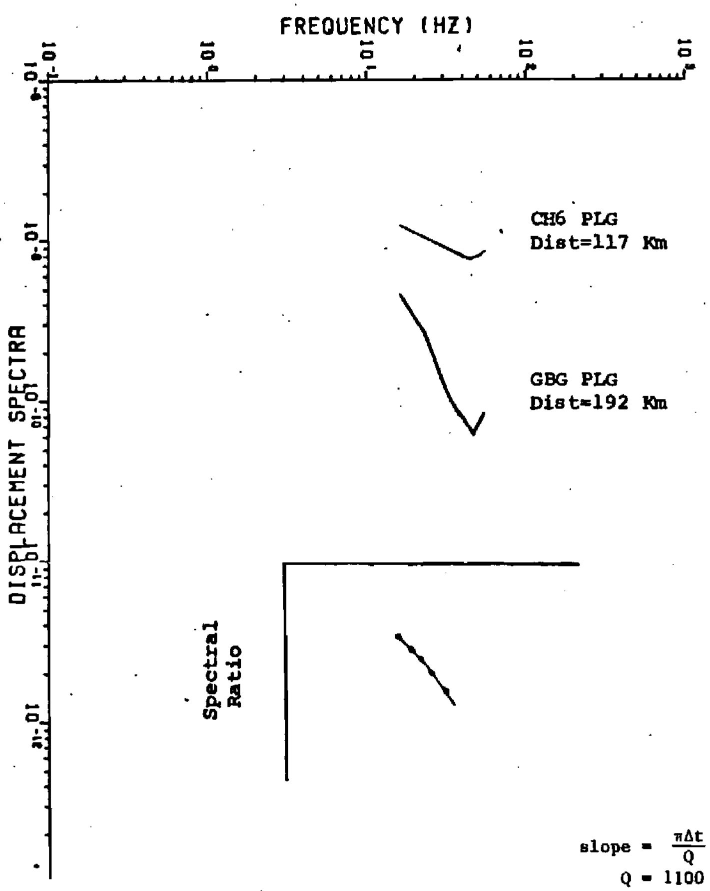

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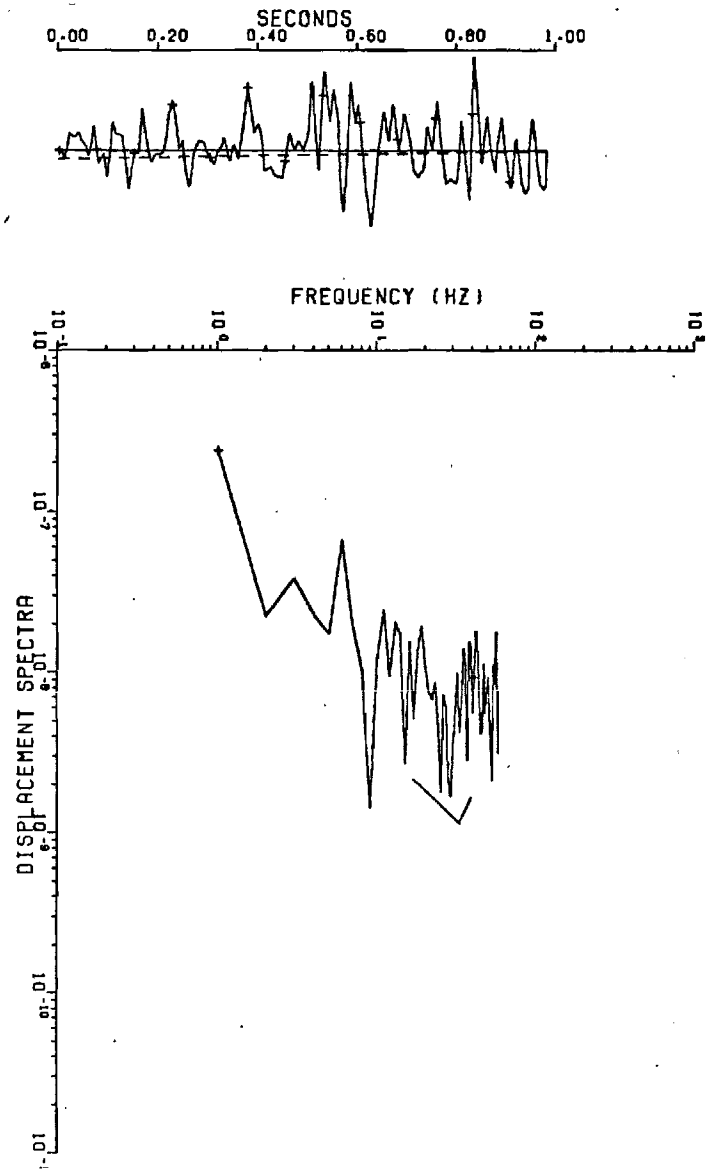

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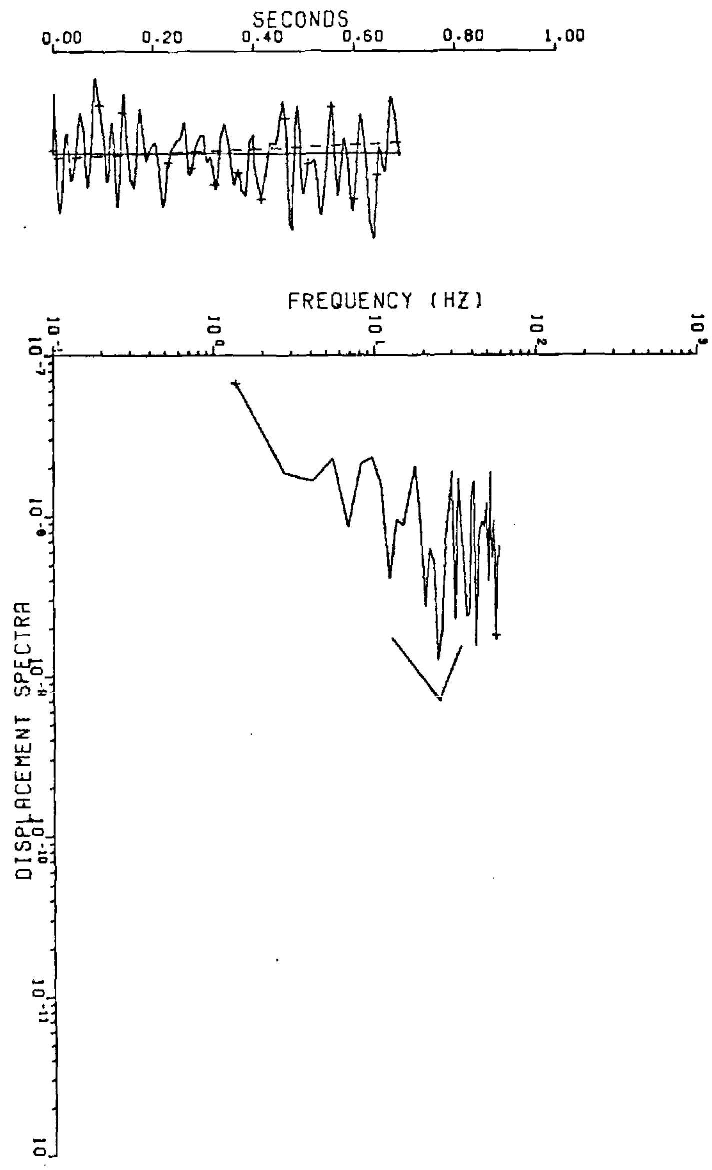

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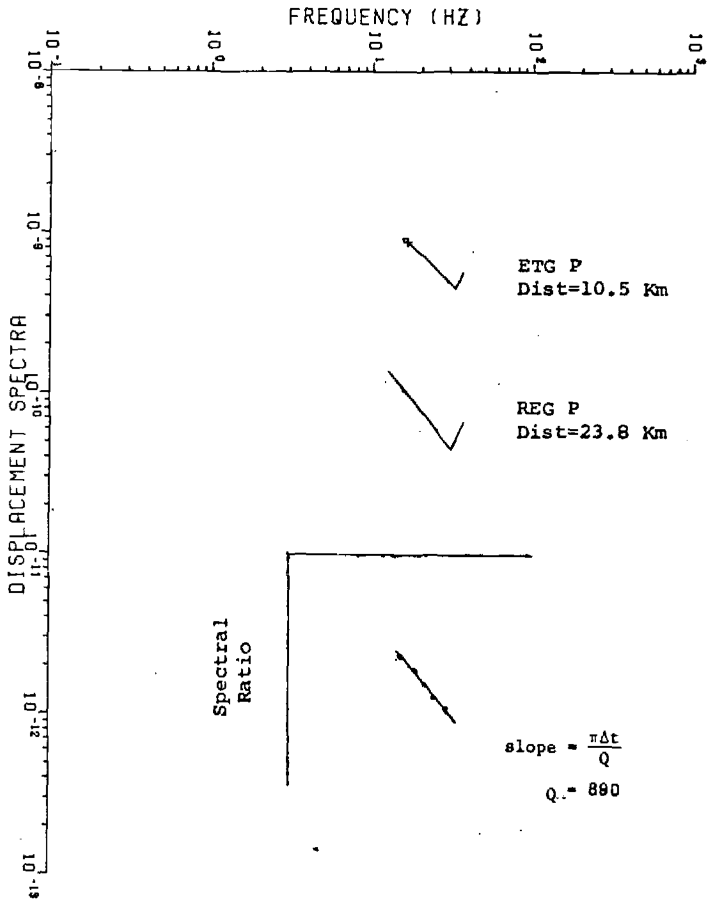
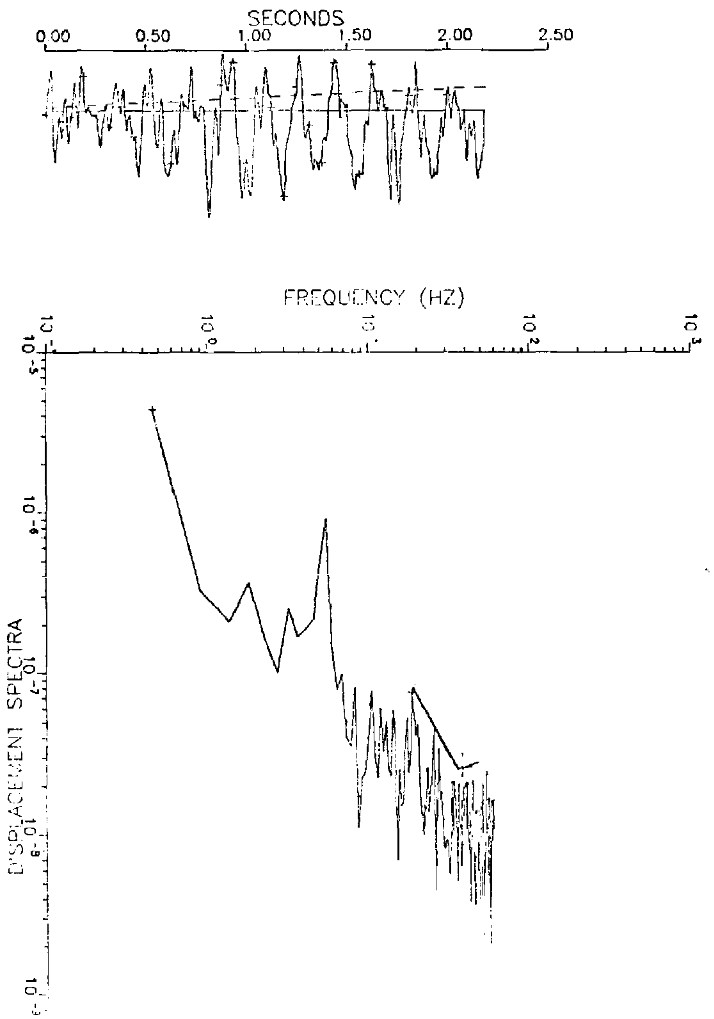
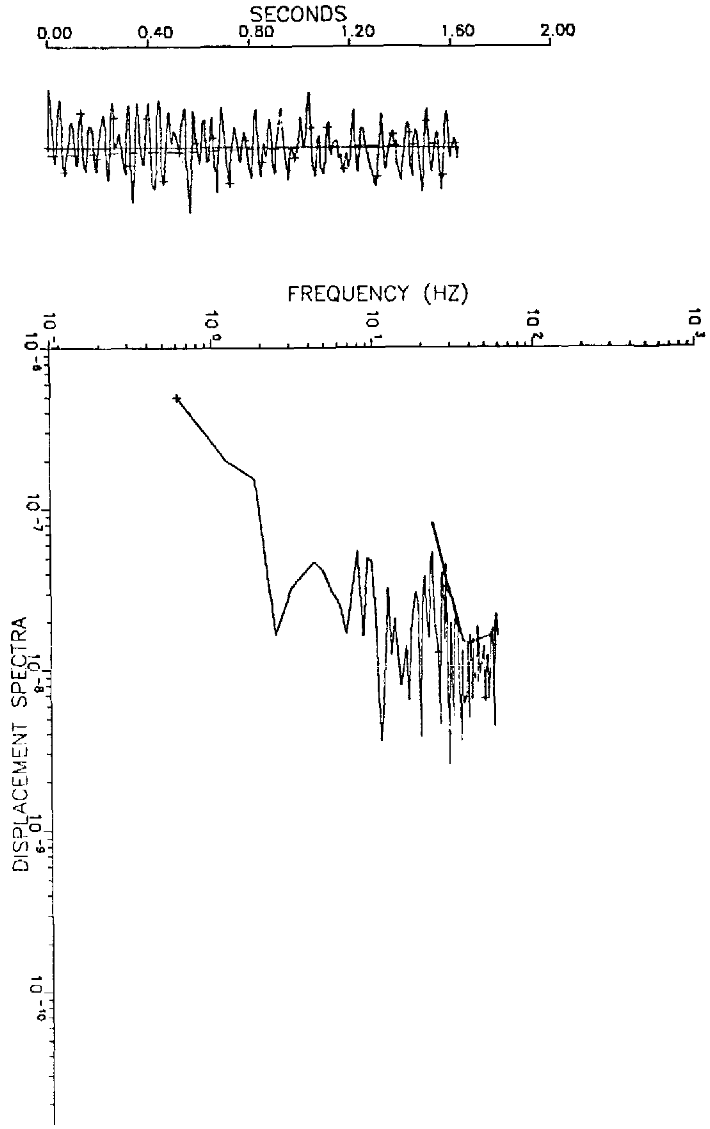


LSA 5/2/78 WDG P



LSA 5/2/78 CH5 P



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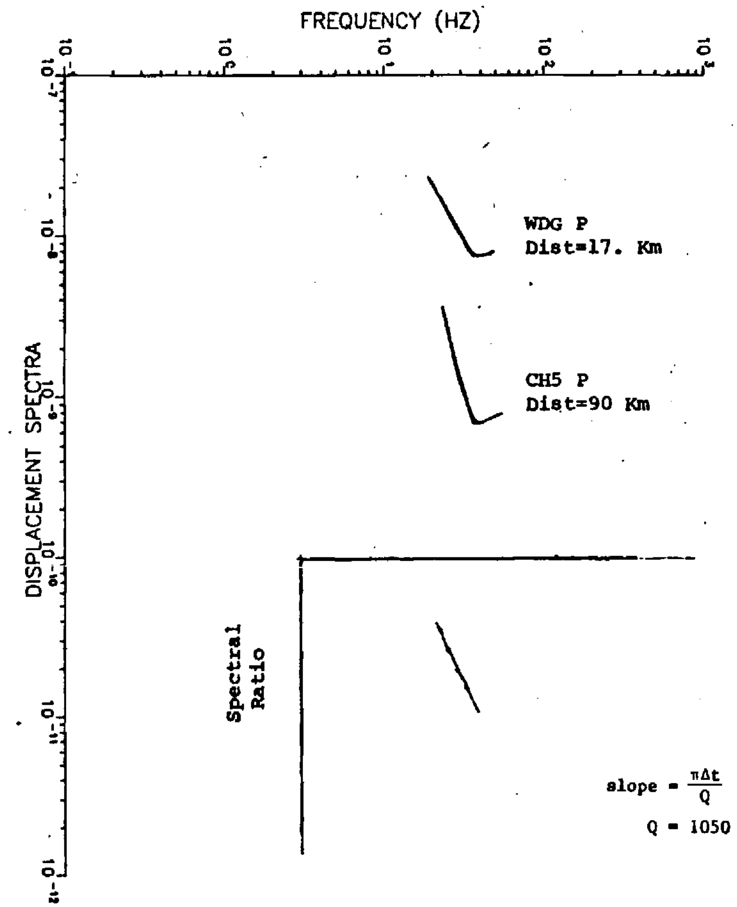
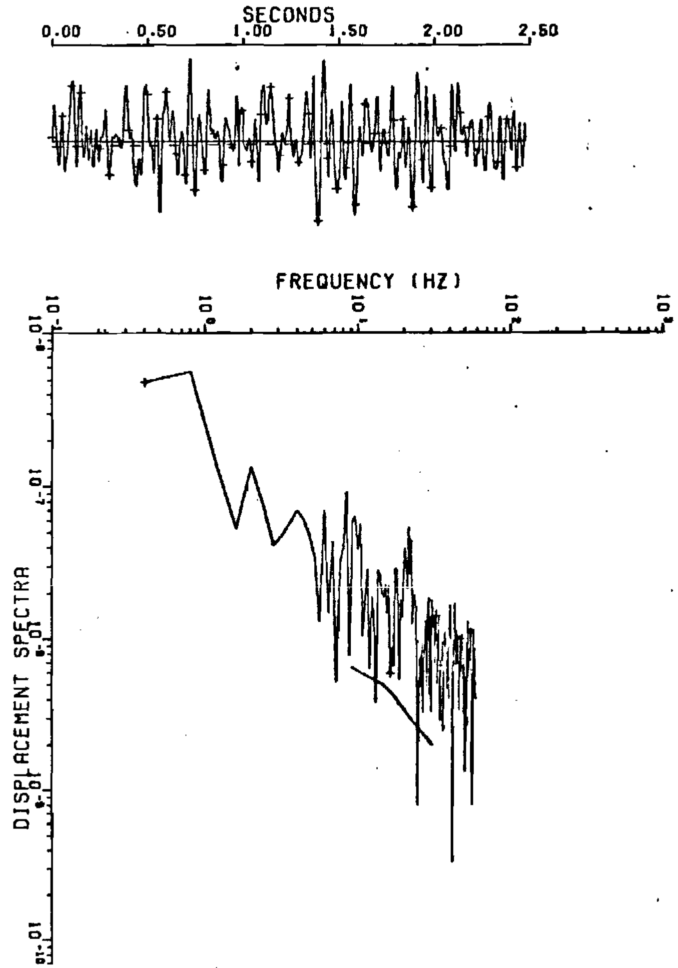
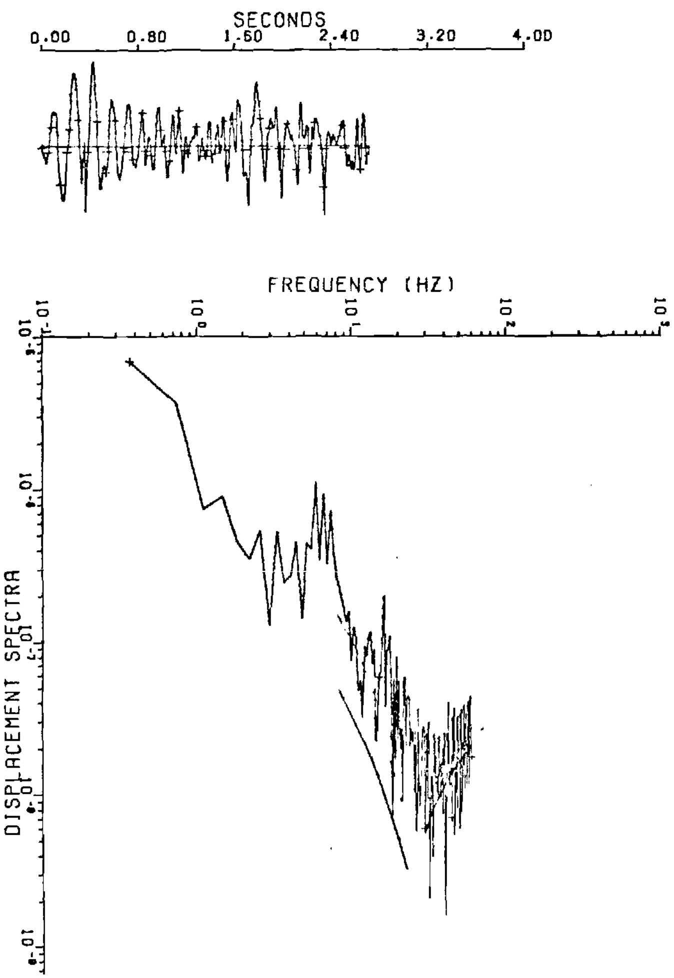


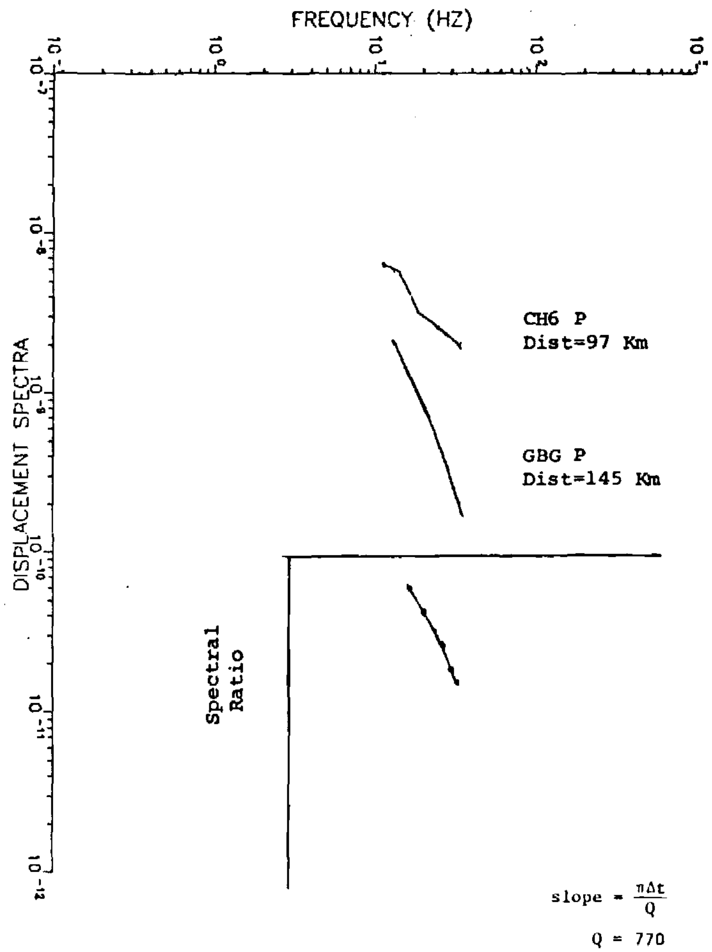
CH6-P L. KEOWEE 1/19/79 MB=2.0



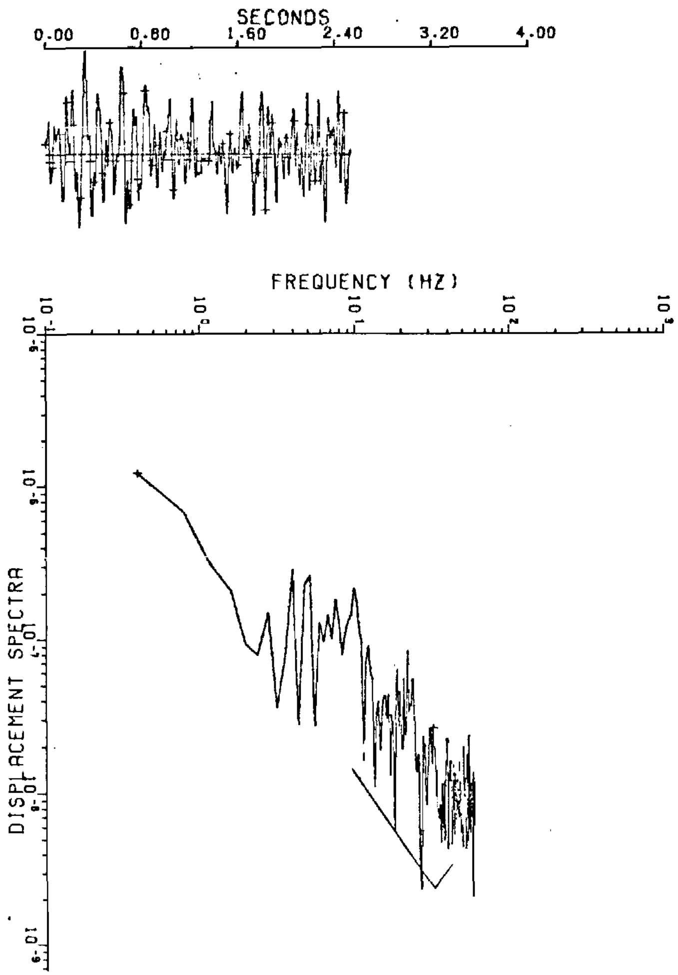
GBG-P L. KEOWEE 1/19/79



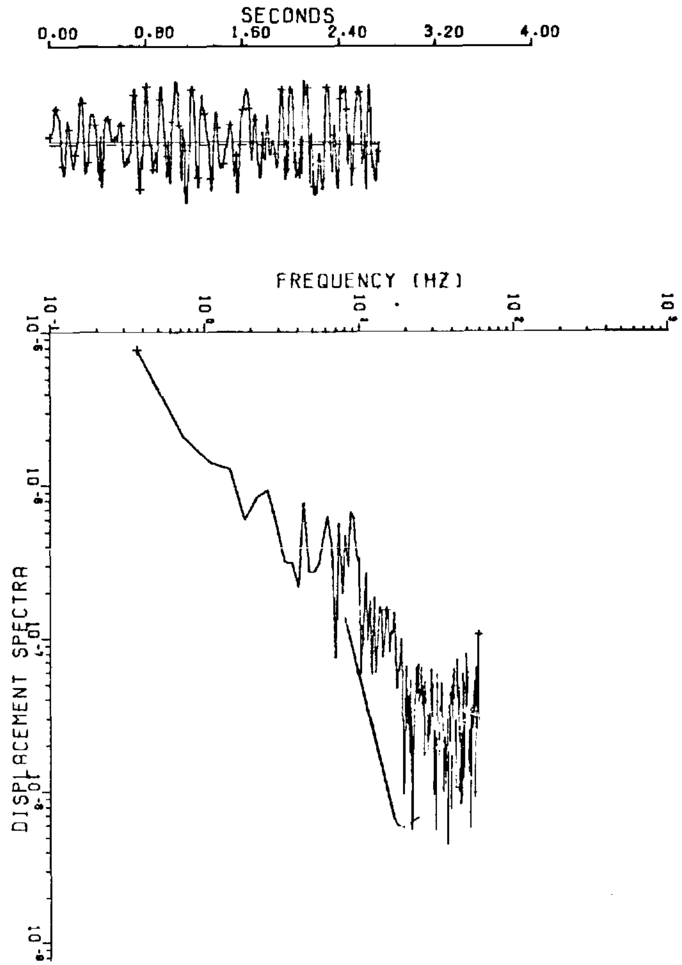
LAKE KEOWEE 1/19/79 M=2.0



CH6-S L. KEOWEE 1/19/79



GBG-S L. KEOWEE 1/19/79



LAKE KEOWEE 1/19/79